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SHORT CONTRIBUTIONS TO CALIFORNIA GEOLOGY

GEOLOGIC HAZARDS

The Ground Shaking Hazard, An Important Consideration in Land-Use Planning. By Walter W. Hays 1

California Earthquakes of April 19–29, 1892. By Diana C. Dale 9

STRUCTURAL GEOLOGY

Thrust Plates of the North-Central Klamath Mountains near Happy Camp, California. By Christopher W. Klein 23

Limitations on Displacement and Southeastward Extent of the Death Valley Fault Zone, California. By Gregory A. Davis..... 27

Tectonic Setting of the Modoc Region, Northeastern California. By Judith L. Hannah..... 35

San Juan-St. Francis Fault—Hypothesized Major Middle Tertiary Right-Lateral Fault in Central and Southern California. By Drew P. Smith 41

PETROLOGY

Syntectonic Batholithic Rocks in South-Central San Diego County, A Preliminary Report. By Victoria R. Todd..... 51

Petrography and Metamorphism of the Morena Reservoir Roof Pendant, Southern California. By Richard G. Berggreen and Michael J. Walawender 61

The Preston Peak Ophiolite, Klamath Mountains, California, An Ancient Immature Island Arc: Petrochemical Evidence. By Arthur W. Snoke, H. R. Bowman, and A. J. Hebert..... 67

STRATIGRAPHY

Middle Eocene Molluscan Assemblage and Stratigraphy, Lower Piru Creek, Transverse Ranges, California. By Richard L. Squires 81

Late Proterozoic-Early Cambrian Biostratigraphy, Correlation, and Paleo-environments, White-Inyo Facies, California-Nevada. By Gail J. Wiggett..... 87

Late Precambrian Depositional Environment of the Pahrump Group, Panamint Mountains, California. By Theodore C. Labotka and Arden L. Albee 93

MARINE GEOLOGY

Clay Mineralogy of Recent Marine Sediments Found on the Southern California Outer Continental Shelf. By Richard W. Berry and Bruce Nocita 101

PREFACE

This Special Report is a compilation of recently completed original geologic investigations, which emphasizes the great diversity and complexity of geologic problems in California. The articles deal with areas ranging from the northern Klamath Mountains to San Diego, and from the desert to the sea.

The thirteen articles in this report are organized into five sections: Geologic Hazards, Structural Geology, Petrology, Stratigraphy, and Marine Geology. Release of this report was timed to coincide with the 1977 annual meeting of the Cordilleran Section of the Geological Society of America, at Sacramento State University, April 5 to 7, 1977.

This report was released about twelve months from the time the call for papers was first sent out. Authors were invited to submit original articles, not to exceed 12 pages, and reviewed by two colleagues. Individual authors are solely responsible for the validity of data and judgements presented.

The division is grateful for the excellent cooperation, and especially the prompt response, from each of the authors and reviewers.

Thomas E. Gay Jr.
State Geologist

E GROUND SHAKING HAZARD, N IMPORTANT CONSIDERATION IN LAND–USE PLANNING[1]

WALTER W. HAYS[2]

ABSTRACT

e Seismic Safety Element, adopted by the California Legislature in 1971, res consideration of all earthquake hazards. In the total set of earthe hazards, ground shaking in an area is one of the most difficult to ct and to quantify because of the lack of knowledge about the thickness physical properties of the near–surface geologic materials. The 1971 San nda earthquake and aftershock sequence provided an opportunity to ire empirical data about the relative ground shaking in the Glendale, rnia, area. An eleven–station seismic array of L–7 velocity seismans was deployed following the earthquake. All of the seismographs in rray, except two, Griffith Park Observatory and Brand Park fire road, located on alluvium of varying thickness. Two of the stations, Glendale cipal Building and Griffith Park Observatory, also recorded the main k. Estimates of the horizontal component, main shock, velocity response ra were produced for the nine sites in the array that did not record the Fernando earthquake. Comparison of these spectral estimates and the ive transfer functions between each site and Griffith Park Observatory ated on rock) showed that frequency–dependent amplification occurred e Glendale area. The relative ground shaking levels correlated fairly well the limited knowledge about the characteristics of the near–surface ogy in the area and provided information that can be used in the –use planning process.

INTRODUCTION

The objective of this paper is to present empirical data recordn the Glendale, California, area from the 1971 San Fernando hquake and aftershock sequence. These data depict the rela ground shaking in the Glendale area and provide informa that can be used in the land–use planning process.

Urban planners and public officials in California and in other ts of the United States have become increasingly concerned ut earthquake hazards. In the past several years, a great deal ffort in California has been directed toward the development implementation of procedures for minimizing damage and ucing loss of life in the event of an earthquake. In 1971, the ifornia legislature adopted an amendment to the State Plang Law which included a Seismic Safety Element as a manday element of the General Plan (see California Council on ergovernmental Relations, 1972 and 1973). This requirement, ng with concerns for other geologic hazards and conserion of natural resources, has focused the attention of city nners and officials on the contributions that geology and er earth sciences can make to the planning process (Mader, 72; Hays and others, 1974; Nichols and Buchanan–Banks, 74). In 1974, the Seismic Safety Commission was created by California Legislature. The present purpose of this commisn is to strengthen earthquake safety in the state by improving blic policy as it relates to reducing earthquake hazards.

Of the various earthquake hazards considered in the Seismic fety Element, earthquake–generated ground shaking, is one of most difficult seismic hazards to predict and to quantify. It s been recognized and widely documented in the literature (e.g., Sezawa and Kanai, 1930; Seed and Idriss, 1969; Murphy and others, 1971a; Shannon and Wilson, Inc., and Agbabian Associates, 1972) since the early 1900's that the soil properties at the recording site can significantly modify the amplitude level and spectral composition of the ground motion recorded at the surface on soil relative to that recorded at nearby sites on bedrock or firmer materials. Modification of the input motion characteristics by the soil column of a particular recording site causes the amplitude of the surface ground motion to be increased in some range of frequencies and decreased in others. The amplitude of the amplified motion is a function of the contrast in physical properties (shear wave velocity, density, damping) and the geometry of the interface between the soil and underlying rock. The frequency range that is affected is a function of the thickness and physical properties of the soil column.

The seismic response phenomenon is complicated by the fact that soils behave nonlinearly under high–strain seismic loads (Seed and Idriss, 1969; Joyner and Chen, 1975). Although there is some controversy about modeling nonlinear soil behavior, it is generally agreed that soil nonlinearities and inelasticity become critically important under certain conditions and may attenuate, rather than amplify certain frequency components in the surface motions relative to the bedrock motions.

CHARACTERISTICS OF THE SEISMIC ARRAY IN GLENDALE

Glendale is located about 30 km southeast of the epicenter of the San Fernando earthquake. An accelerogram from the main shock was recorded at two locations in the area: (1) at the Glendale Municipal Building, 633 East Broadway (Station 41) and (2) at the Los Angeles Griffith Park Observatory (Station 46) (see figure 1). The levels of peak horizontal acceleration were, respectively, 28 percent and 18 percent g. In addition, aftershocks were recorded on an eleven–station seismic array (table 1) deployed in the area (figure 1). The aftershocks were recorded on the L–7 velocity seismograph (King, 1969), a portable, three–component system that records particle velocity on magnetic tape and has a flat response from 0.1 to 34 Hz. The high accuracy (± 5 percent) to which the L–7 system can be calibrated and the broad–band frequency response make the system very useful in engineering seismology research (Park and Hays, 1976).

Table 1. Stations in the Glendale area used to record aftershocks.

Station	*Code*
Brand Park fire road	33
Casa Adobe de San Rafael	34
Columbus School	35
Edison School	36
Verdugo Park	38
Nibley Park	39
Woodrow Wilson Junior High School	40
Glendale Municipal Building*	41
Glendale High School	42
Griffith Manor Park	45
Griffith Park Observatory*	46

*Station recorded main shock and aftershocks.

viewed by Albert M. Rogers and Richard E. Warrick.
S. Geological Survey, Denver Federal Center, Denver, Colorado 25

SURFICIAL GEOLOGY IN THE GLENDALE AREA

All of the L–7 seismographs in the array except two were located on alluvium. The stations at Brand Park fire road (33) and Griffith Park Observatory (46) were located on granitic rock. The alluvial soil throughout the Glendale area was derived from the parent bedrock of the bounding Verdugo and Santa Monica Mountains, located respectively on the north and west of the area (Lamar, 1970). The thickness and physical properties of the alluvium deposits throughout the Glendale area are not well established. On the basis of published and unpublished data which includes detailed field mapping at selected sites in the Glendale area, the alluvium tends to be thin along the base of the mountains, as is shown schematically in figure 1. The alluvium ranges in composition and consistency from very fine–grained clays to coarse–grained sand, gravel and cobble sediments. The physical properties of the alluvium and underlying rock are not well known except at several specific locations where drilling logs were available or where refraction surveys were made (Woodward–McNeill and Associates, 1973; Duke and others, 1971). These limited data served to define a reasonable range for the shear wave velocities and thicknesses of the soil columns used in this study.

SAN FERNANDO EARTHQUAKE GROUND MOTION ESTIMATES

Although the San Fernando earthquake produced more strong motion accelerograms than any prior earthquake, the ground motion data sample was insufficient to quantify the effect of near–surface geology on the ground motion experienced within the Glendale area. The aftershock data recorded on the seismic array deployed in Glendale area, however, complement the strong motion data sample (Hays and others, 1975a) and provide a basis for estimating the relative ground response caused by the San Fernando earthquake (Hays, 1976). Estimates of the horizontal component, main shock pseudo relative velocity (PSRV), response spectra were produced by Murphy and others (1971b), using data from six aftershocks recorded simultaneously on the array. The response spectrum depicts the maximum amplitudes of response of an ensemble or simple, damped oscillators to the input ground motion, and is a model used by structural engineers. The variability within the main shock spectral estimates is shown in figure 2. In this figure, upper and lower envelopes constructed from horizontal component spectral velocity estimates for the nine stations of the array that did not record the San Fernando earthquake are compared with the actual response spectra derived from the main shock accelerograms recorded, respectively, at Glendale Municipal Building (41) and Griffith Park Observatory (46).

The relative transfer function between two sites at similar distance and azimuth from the epicenter provides information about frequency–dependent effects caused primarily by the differences in the near–surface geology. Figure 3 shows the average horizontal component transfer functions for: (1) Verdugo Park (Station 38) relative to Griffith Park Observatory (Station 46), (2) Verdugo Park relative to Glendale Municipal Building (station 41), and (3) Glendale Municipal Building relative to Griffith Park Observatory. Station 38 is located on thin alluvium near the base of the Verdugo Mountains (see figure 1); Station 41 is located on thick alluvium; and Station 46 is located on granitic rock. The first two relative transfer functions show a significant increase in the amplitude of ground response at the 0.2 second period. The empirical transfer function for thin alluvium/thick alluvium shows substantial amplification for perio of less than 0.3 second and essentially identical response periods greater than 0.4 second. Thus, the short–period (hig frequency) components are attenuated more in the thick secti of alluvium than in thin section of alluvium. The empirical tra fer function for thin alluvium/rock shows a generally high amplitude for all periods, except in the vicinity of 0.5 second a above 2.0 second, where the responses are essentially equal. T transfer function for thick alluvium/rock does not show ampli cation at short periods that the thin alluvium/rock transfer fu tion shows.

The empirical transfer functions are similar to those th would be predicted theoretically for the response of a single la system to incident SH waves. Using as a guide the informati about the shear wave velocities and layer thicknesses reported Duke and others (1971), plausible geologic models can be co structed which agree with the relative site transfer functio Figure 4 shows one such geologic model which assumes: (1) 29 m thick alluvium layer at Station 38; (2) a 62 m thick all vium layer at Station 41; and (3) shear wave velocities (β) 500 m/sec and 1,900 m/sec, and densities (ρ) of 1.5 g/cc a 2.0 g/cc respectively, for the alluvium and underlying rock. T theoretical transfer function between the two sites was comput (Murphy and others, 1971b) using the geologic model and viscosity to yield 2.5 percent the critical damping value in t fundamental mode; it is compared with the empirical transf function in figure 4. The agreement is good which suggests th the different ground motion response observed at the two sit can be explained by the difference in thickness of the alluviu

Two map views that show the variation in estimated ma shock ground shaking, relative to that of Griffith Park Observ tory, are shown in figures 5 and 6. The first map shows t smoothed relative ground response for the period band 0.05 0.16 sec. The contours are arbitrarily made to conform with t interpreted soil and rock distribution in the area (see figure 1 The largest relative ground response occurs at Station 38 an generally coincides with the distribution of thin alluvium alon the mountain front, especially in the northeast portion of t area. This is the zone where amplification of the high frequenc ground motions would be expected. The second map depicts th estimated main shock relative response for the 1.82–2.46 secon period band. The largest values of relative ground response a in the center of the area and would be expected to correspon with the thicker alluvium deposits in the area.

DISCUSSION

Ground response maps such as those derived for the Glendal area are important input for the land–use planning proces which considers earthquake hazards. Ground shaking not onl contributes directly to losses through vibratory damage to man-made structures but also can trigger secondary effects such a landslides and other modes of ground failure. Land–use planning is concerned primarily with decisions which affect the compre hensive arrangement of the uses of the land. A planning decisio is basically the process of making a choice between alternativ actions based upon a set of socioeconomic criteria and an esti mate of what the outcome of each planning action might be. Fo example, some criteria are expressed spatially (e.g., physical development constraints) and some are expressed aspatially (e.g., law and order); some quantitatively (e.g., economic feasibility) and some qualitatively (e.g., public opinion). The objective of the planning process in earthquake prone areas is to find the most optimum locations for land uses when all the planning, socioeconomic, and earthquake hazards criteria are taken into consideration.

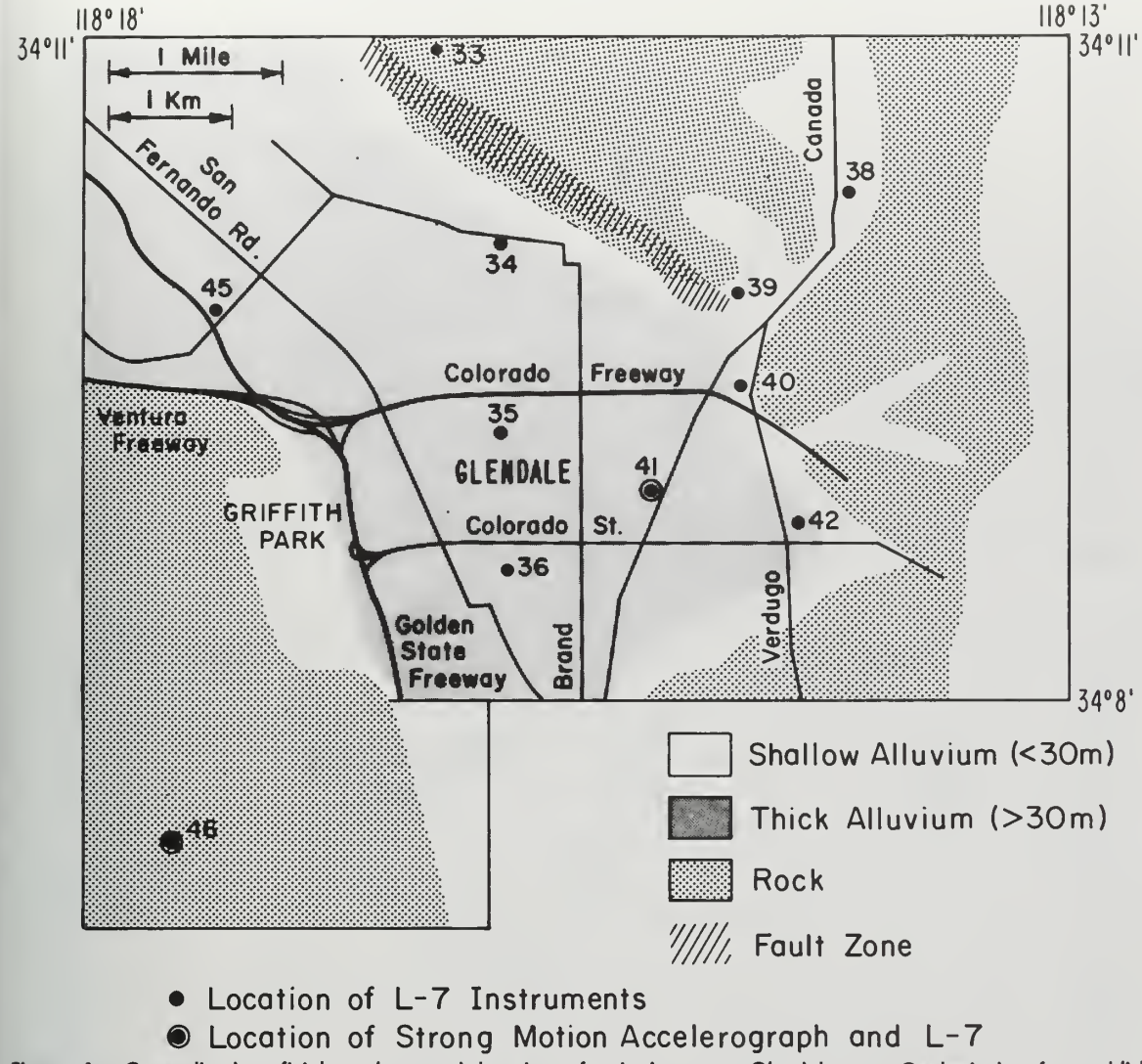


Figure 1. Generalized surficiol geology and location of seismic orroy, Glendale oreo. Geologic data from published (Lamar, 1970; Duke ond others, 1971) ond unpublished sources.

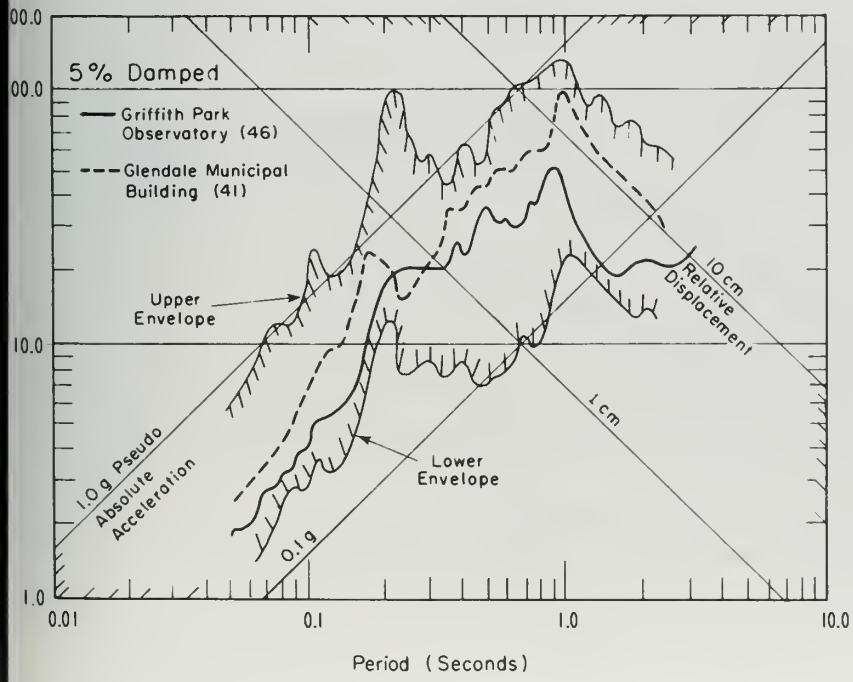


Figure 2. Comporison of upper and lower envelopes of estimoted moin shock, horizontal component response spectra with octuol spectra derived for the Griffith Park (rock) ond Glendole Municipol Building (thick alluvium) sites.

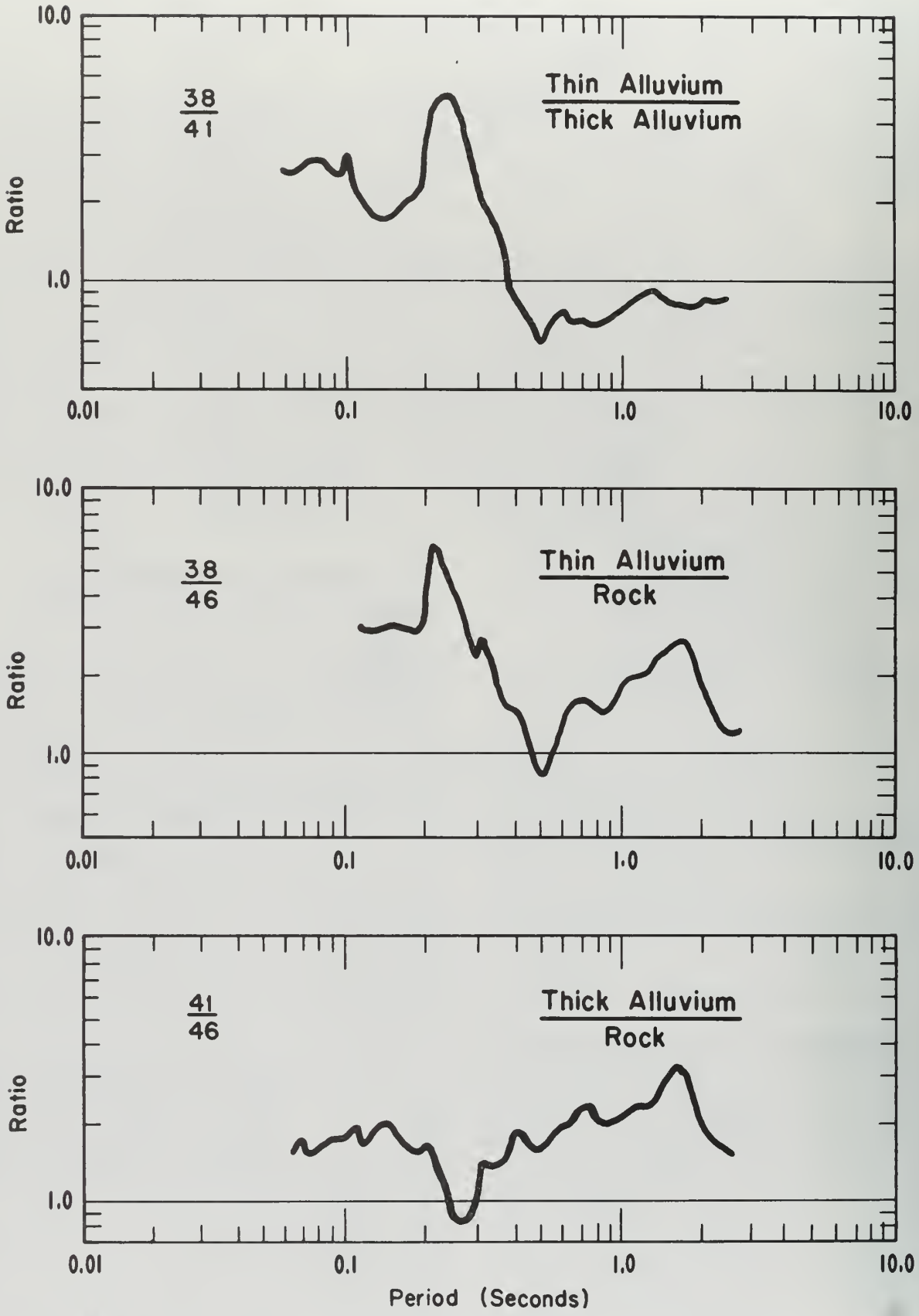
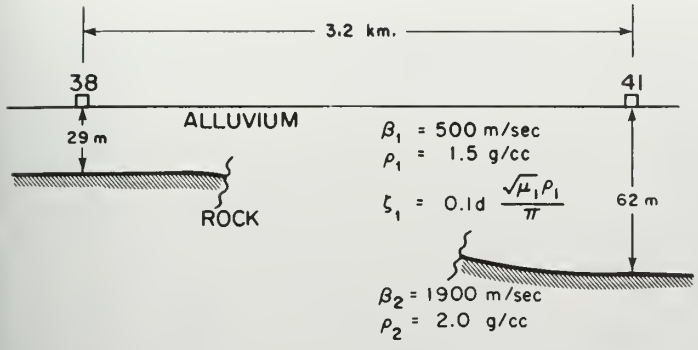


Figure 3. Average horizontal component transfer functions for rock, thin alluvium, and thick alluvium sites.



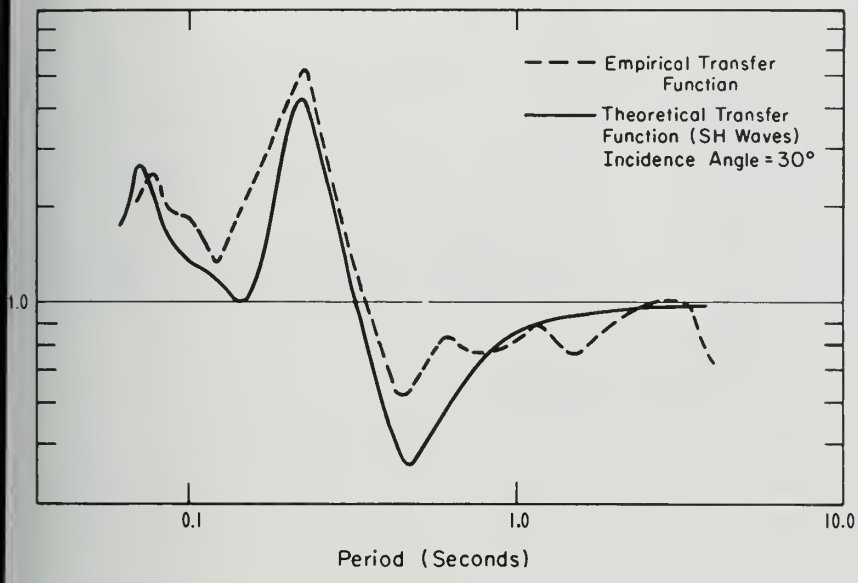
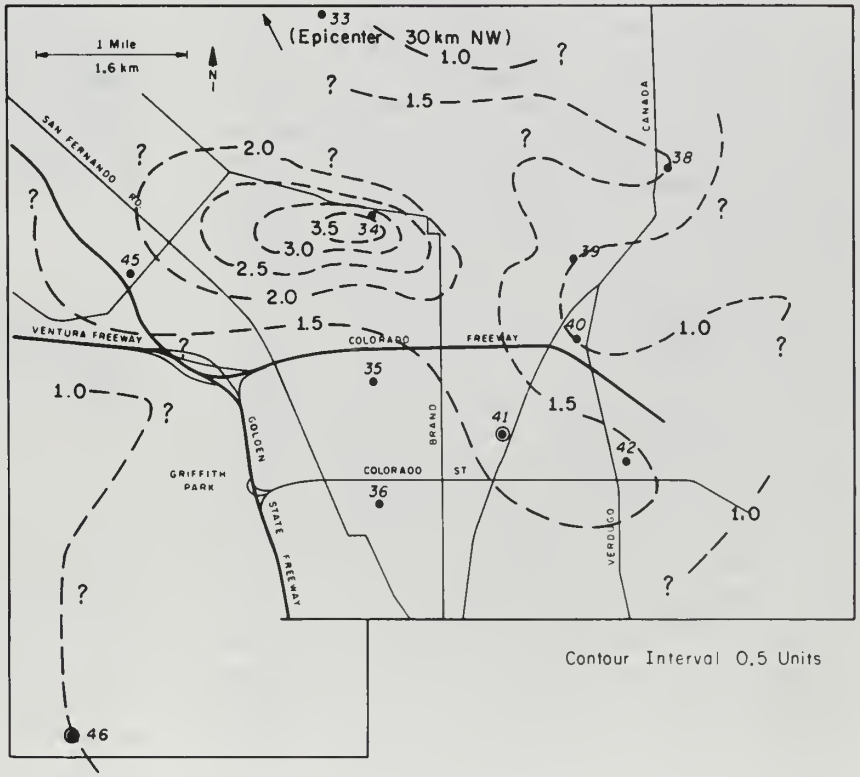


Figure 4. Generalized geologic model (top) and comparison of theoretical and empirical transfer functions (bottom) for Stations 38 and 41 (thin alluvium/thick alluvium.



ure 5. Generalized contour map of es-ated relative horizontal ground response 971 San Fernando earthquake, 0.055–6 second period band.

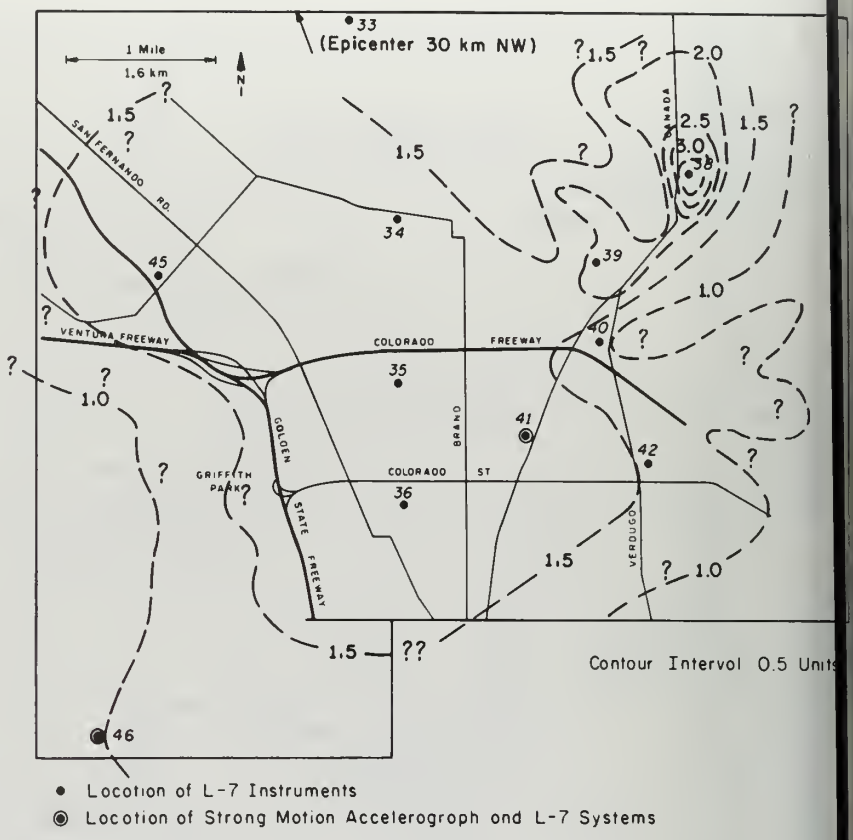


Figure 6. Generolized contour map of estimoted relotive horizontol ground response in 1971 San Fernando earthquoke, 1.82–2.46 second period band.

Ground response maps help to delineate those areas of Glendale for which the frequency–dependent effects of ground shaking are expected to be least, intermediate, and greatest during a future earthquake. This information can be incorporated quantitatively in the planning decision process because different types of land–use have different sensitivities to ground shaking. For example, the sensitivity of a hospital's exposure to ground shaking of a certain level is quite different from the sensitivity of a warehouse's exposure to the same ground shaking. Good land–use planning also identifies and encourages favorable site–structure combinations. Because man–made structures with natural periods coinciding with those of the underlying soil deposits have a greater susceptibility to damage (all other factors being equal) than structures with natural periods that differ from the underlying soil deposits, certain classes of structures might be evaluated as being unsuitable for certain ground response zones in Glendale. For example, low–rise, acceleration–sensitive buildings might be better located, from a ground shaking view, at sites in Glendale other than along the Verdugo mountain front where thin alluvium overlying rock causes the high frequencies and ground acceleration to be amplified. Also, high–rise buildings with a fundamental mode of response in the 1.8–2.5 second period range might be better located at sites other than in the center of the area where thick alluvium deposits cause amplification of the long period seismic waves. In the planning process, many other criteria (e.g., other earthquake hazards such as surface faulting, landslide susceptibility, liquefaction potential, inundation potential; and structural design considerations) would be incorporated with ground shaking and other criteria to develop a workable land–use plan for the area.

A number of investigators (e.g., Wentworth and Yerkes, 19 Farhoomand and Scholl, 1972; Hays and others, 1975b; Ste brugge and others, 1975; Borcherdt, 1975) have reported various facets of research in the complex earthquake hazar reduction problem. These results are encouraging and offer ho that improved land–use planning relative to earthquake hazar will become a practical possibility in the near future.

ACKNOWLEDGEMENT

The helpful suggestions made by Dr. Albert M. Rogers a Mr. Richard E. Warrick, Branch of Earthquake Hazards, U Geological Survey, in technical review are gratefully ackno edged.

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LIFORNIA EARTHQUAKES OF APRIL 19–29, 1892[1]

DIANA C. DALE[2]

ABSTRACT

etween April 19 and 29, 1892, two major earthquakes and numerous rshocks produced massive destruction in the towns of Winters, Vacaville, Dixon at the west edge of the Central Valley and felt-effects of varying ree from the Pacific coast eastward into Nevada. Isoseismal maps have n derived for the large events of April 19 and 21 and a felt-area map the aftershock of April 29, 1892. The isoseismals display a NNW–SSE ngatian, particularly in the areas of high intensity. The events of April 19 21 produced felt-areas of slightly more than 90,000 sq. miles. The April arthquake had a maximum intensity of VIII–IX (Modified Mercalli) and mated Richter magnitude of 6.7. The event of April 21 was slightly nger, with a maximum intensity of at least IX (MM) and corresponding gnitude of about 7.0. Epicenters of the events were located in the hills t of the Central Valley. The April 19 earthquake was centered between, d to the west of, Vacaville and Winters. On April 21, the epicentral region s to the north, in the ridge area west of Winters. The April 29, 1892, rshock, of maximum intensity V–VI (MM), may have originated somet south and west of the two major shocks.

INTRODUCTION AND METHOD

During the period of April 19 through 29, 1892, a series of thquakes originating east of the San Francisco Bay was felt oughout central California from the coast into Nevada. Two hese events, occurring on April 19 and 21, produced extensive mage to towns along the western margin of the Central Valley.

Although a great deal of information about these earthquakes vailable, principally from contemporary newspaper coverage, omprehensive analysis of the April 19 and 21, 1892, earthakes had never been performed.

All previous descriptions of these events have been based on work of Charles D. Perrine. The information appearing in his arthquakes in California in 1892" (1893) was copied in E. S. iden's Catalogue of 1898 and was utilized in the later works Harry Fielding Reid, Harry O. Wood (1916), and Townley d Allen (1939). The descriptions of these earthquakes in arthquake History of the United States" (1973) are likewise ived from Perrine and Holden.

"Earthquake History of the United States" associates the ril 19, 1892, event with Vacaville and that of April 21 with nters and gives a maximum intensity of IX (MM) for both nts. More detailed information than that contained in the rces heretofore used as authorities is needed in order to furr interpret these events.

Such additional definition has recently become of particular concern in the effort to relate the major seismic events of California to geologic structures and ultimately to guide the design of public utilities and other projects within central California. Accordingly, an investigation of these earthquakes was undertaken by the author in the Spring of 1975, with particular emphasis on establishing the location of the epicenters.

The basic method utilized in this investigation was the collection and analysis of original newspaper and lighthouse reports from the time of the events. Some 40 newspapers were scanned and items describing the earthquakes studied in groups according to location affected. Intensity, duration, and general nature of shaking were determined for each location; a series of isoseismal maps were produced; and estimates of total felt area and magnitude were calculated.

This paper outlines the analysis of the two major earthquakes of April 19 and 21, 1892, and the smaller event of April 29. Although a number of minor shocks were also recorded throughout the 11-day period, the reports are vague or incomplete, and therefore no study has been performed on these small events.

APRIL 19, 1892, EARTHQUAKE

Intensity Data, Isoseismal Map, and Felt Area

Descriptions of the April 19, 1892, earthquake have been recovered by the author for 99 inland locations and nine lighthouses. This number includes 47 towns which had not been considered by the earlier works concerned with this earthquake. Table 1 lists the intensities assigned to each place from which information has been collected.

An isoseismal map (figure 1) for the April 19, 1892, earthquake has been constructed from the data presented on table 1. The farthest points in each direction reporting the earthquake were Redding to the north, Bridgeport to the east, Fresno to the south, and Half Moon Bay and Santa Cruz to the west. The total felt area for the event, based on this information, was approximately 92,000 sq. miles.

The isoseismal pattern generated by this earthquake had a N–S elongation, the area of higher intensities (VIII–IX and VII MM) was oriented NNW–SSE, while the lower intensity areas (VI, V, and IV) appear to have been oriented more nearly NW–SE. Attenuation was fairly rapid in the areas of high intensities and toward the western coastal region, but occurred much more slowly to the east, where intensity V and IV (MM) effects were produced over a large area of the Central Valley and Sierra Nevada.

viewed by Cole R. McClure, Marilyn Trouwborst, and Joe Litehiser. chtel Incorporated, San Francisco, California.

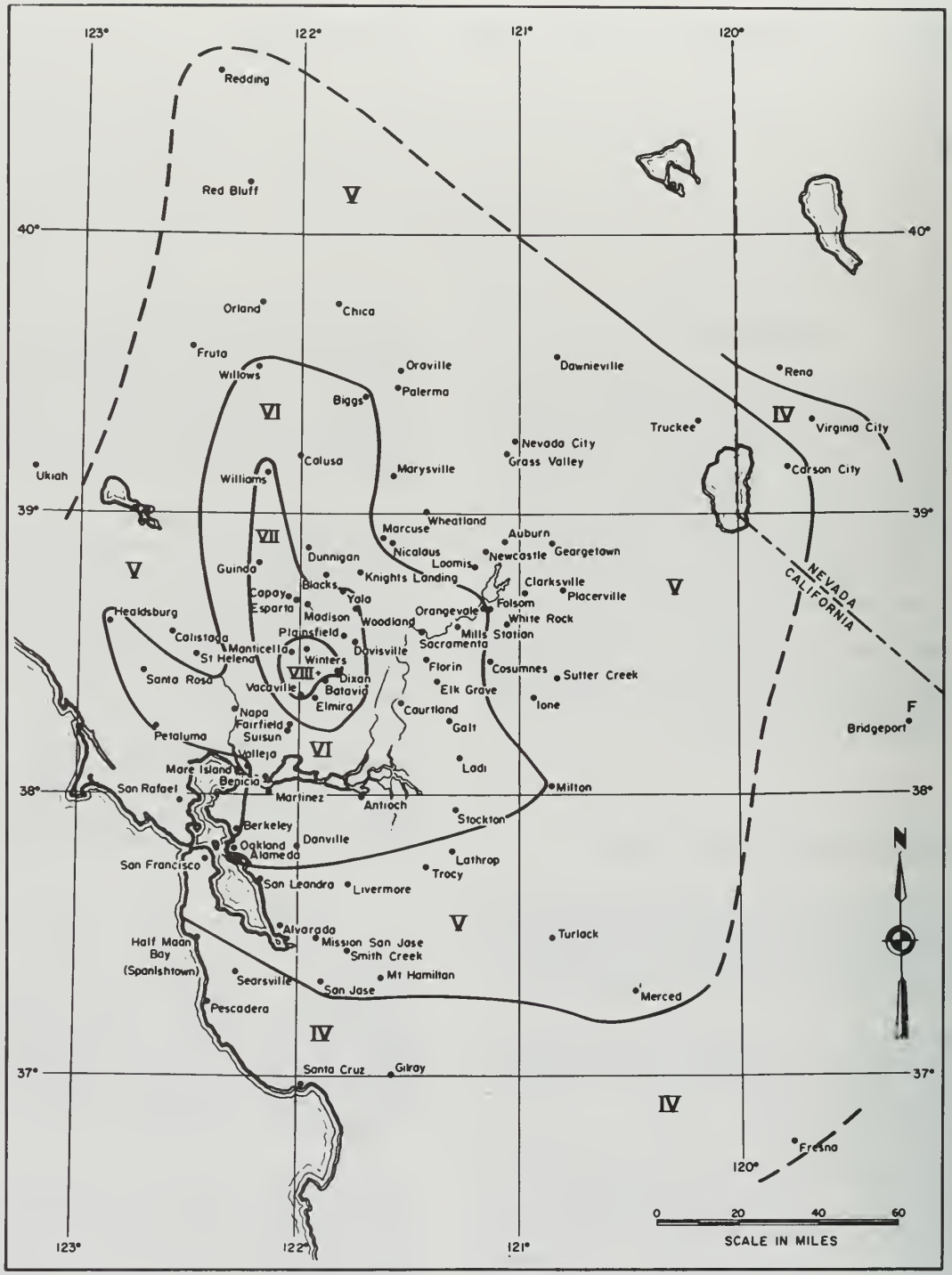

Figure I

ISOSEISMALS OF APRIL 19, 1892 EARTHQUAKE

Table 1

APRIL 19, 1892 EARTHQUAKE INTENSITIES

Town	Intensity
eda	?
rado[a]	V
och	VI
rn	IV-V[b]
via[a]	VII
cia	VI
eley	V?
s	V-VI
ks[a]	VI
geport	?
stoga[a]	V?
y	VII-VIII
on City, Nev.	V
o	low V or IV-V
rksville[a]	V
sa	V-VI
rtland[a]	V-VI?
ville[a]	VI-VII?
isville	VI-VII
n	low VIII
nieville[a]	V?
nigan[a]	VI
Grove[a]	V-VI
ira	VII
arto	VII-VIII
rfield	VI-VII, VII
rin[a]	V-VI?
som	V-VI
sno	IV
to[a]	not felt
t[a]	V-VI?
rgetown[a]	?
roy[a]	IV?
ss Valley	V(IV-V)
nda[a]	VI?
f Moon Bay[a]	IV or V[c]
ldsburg	VI
e[a]	V?

Town	Intensity
Knights Landing[a]	VI
Lathrop[a]	V?
Lobitos[a]	IV
Lodi[a]	V-VI?
Loomis[a]	V
Madison[a]	VI
Marcuse	V?
Mare Island[a]	V-VI
Martinez	VI
Marysville	V
Merced	V
Mills Station[a]	Heavy
Milton	V-VI
Mission San Jose[a]	Quite severe
Monticello	?
Mount Hamilton	IV-V?
Napa	V-VI?
Nevada City	V-VI?
Newcastle[a]	IV or V?
Nicolaus	V?
Oakland	VI
Orange Vale[a]	VI
Orland	V?
Oroville[a]	V
Palermo[a]	IV-V?
Pescadero[a]	IV?
Petaluma	V-VI
Placerville	IV?
Plainsfield[a] (between Napa & Winters)	VI-VII
Pleasant Valley	VII & VIII[d]
Red Bluff[a]	V?
Redding[a]	V?
Reno, Nev.	III?
Sacramento	VI
St. Helena[a]	V
San Francisco	V-VI
San Jose	IV-V

Town	Intensity
San Leandro[a]	V?
San Rafael	V
Santa Cruz[a]	IV?
Santa Rosa	VI
Searsville[a]	Heavy
Smith Creek	Felt
Stockton	V-VI
Suisun	VI
Sutter Creek[a]	V
Tracy[a]	V?
Truckee[a]	V
Turlock[a]	?
Ukiah	Not felt
Vacaville	VIII-IX
Vallejo	V
Virginia City, Nev.	IV
Wheatland	V
White Rock[a]	III-IV?
Williams[a]	VII
Willows	VII-VIII
Winters	VIII-IX
Woodland	VII
Yolo[a]	VI-VII
Yuba City[a]	VI
Light Houses	
Angel Island L.H.	III?
Bonita Point L.H.	Sharp
East Brother L.H.	Not heavy
Fort Point L.H.	IV
Lime Point L.H.	IV
Mare Island L.H.	V
Oakland Harbor L.H.	VI?
Roe Island L.H.	VI?
Yerba Buena, L.H.	VI?

[a] Towns not reported by Perrine (1893) and Holden (1898)
[b] IV-V indicates an intensity greater than IV and less than V
[c] IV or V indicates conflicting reports of intensity
[d] VII & VIII indicates a large area which experienced a range of intensities.

Epicenter and Maximum Intensity

Two types of information have been considered in determining the location of the epicenter: (1) cultural, including structural effects and damage to buildings, and (2) reports of natural phenomena such as fissures and landslides.

The towns which sustained the greatest structural damage were Vacaville, Winters, and Dixon. All three towns reported similar types of destruction: wooden buildings twisted or wrenched off their foundations and bulging or collapsed brick walls, firewalls, and roofs. According to newspaper articles and drawings, both Vacaville and Winters reported a similar degree of wreckage; Dixon reported fewer specifically named damaged or destroyed buildings. In terms of dollar loss, though, all of the contemporary reports agreed that Vacaville had experienced the greatest damage. As expressed in the following article,

> "Vocaville [wos] the severest sufferer ... The domoge done is voriously estimoted ot from $100,000 to $150,000. Winters wos the next lorgest sufferer by the shoke, the losses being estimoted ot from $70,000 to $100,000. Dixon wos the next heoviest loser ... The domoge to buildings is estimoted ot from $15,000 to $20,000." (*Grass Valley Union,* April 21, 1892, p. 3)

The difference in dollar estimates of loss between these towns cannot be explained by substantial size differences. Vacaville reported a population of 725 in 1890 and Winters, a popualation of 785 in 1900 (the first census in which Winters was recorded separately). Dixon had a population of 1,082 in 1890.

The comparison of damages led to the widely published judgement that Vacaville was the epicenter of the event. Preceding or following articles listing structural damage are such statements as,

> "Vocoville seemed to be the center of the disturbonce." (*Western Watchman,* April 23, 1892, p. 2)

> "Senotor Rickey of Nevodo wos ot Vocoville, the center of the eorthquoke shock, when it occurred." (*San Francisco Examiner,* April 20, 1892, p. 1)

However, consideration of foundation conditions in Vacaville provide explanation of the greater degree of damage there than elsewhere:

> "Eulotis Creek runs through the town of Vocoville. It wos noticed thot most of the domoge done wos on the south side. This wos moinly becouse neorly oll the buildings ore in thot direction, but it is stoted thot the shock wos not as severe on the north side of the creek. As the wolls of the college ond Christion Church were not domoged the stotement is probobly true. The difference in the violence of the shock is exploined by the chorocter of the soil, which is not so deep on the north side of the creek, offording o solid foundotion." (*Sacramento Evening Bee,* April 20, 1892, p. 1)

Thus, the degree of structural damage in Vacaville may have been due more to foundation conditions than to proximity of the epicenter.

From the above data, it is clear that structural damage alone is an inadequate criterion for locating the epicenter. Of greater value are the numerous articles containing descriptions of natural effects in the countryside around the towns receiving major damage. The following excerpts are representative of the available data:

Winters Vicinity

> "About o mile obove Winters the shock wos so severe thot the bonks, twelve feet in height ond seventy–five yords in length coved into the creek. The woter wos thrown onto the lond o distonce of twenty–two feet beyond the bonks. In onother ploce, there ore fissures in the bed of the creek from one to six inches in width, ond the sound of escoping gos wos distinctly heard this morning." (*Daily Democrat,* April 20, 1892, p. 3)

> "Up the conyon the road is chocked up with boulders dislodged fro hillsides. In one ploce on ocre of ground slid into the creek. In the cree mony fissures coused by the quoke, from some of which gos is hissing f (*Solano Republican,* April 22, 1892, p. 3)

> "On Putoh Creek, holf o mile west of Winters, o phenomenon wos nessed by o young mon nomed Fred Willis, who was riding past ot the of the big shoke. There seemed to be on explosion, ond the woter wos th from the creek to o distonce of twenty feet on either bonk. Then foll o hissing sound os of gos escoping. At doylight severol fissures were f in the bed of the creek ond in the roadwoy ond fields odjoining. On side of the creek where the explosion took ploce the bonks coved in londsides being seventy–five feet in length ond twelve feet deep." *Francisco Examiner,* April 21, 1892, p. 1)

> "Neor the town the bonk of Putoh Creek, ten feet wide, coved in, olong the bottom of the creek for o greot distonce rents were mode b shocks.

> "Up the Berryesso road the passonge is blocked by immense boul some weighing severol tons. They were thrown with mighty force dow hillsides into the road. [It is neor this point where the rents in the road noticed.] From oll oround come reparts of wells filling up with coving ..." (*San Francisco Chronicle,* April 20, 1892, p. 1; *Evening Bee* [with bro ed section above], April 19, 1892, p. 1; *Weekly Delta,* April 21, 1892, *San Jose Mercury* [with brocketed section above], April 20, 1892, p.

Vacaville Vicinity

> "A good many stote thot for some time ofter the shock the oir wos f with sulphurous fumes, which were emitted from the fissures in the gro of which there ore mony hereobouts." (*San Francisco Examiner,* April 1892, p. 1)

Capay–Esparto

> "The eorth opened in severol ploces between here [Esparto] ond Cop (*San Francisco Examiner,* April 20, 1892, p. 1)

These articles indicate that the greatest natural effects of shaking were centered to the north and west of Vacavi between there and Winters. The explosive expulsion of water the landsliding ½–mile west of Winters, the landslides and re up the canyon west of Winters, in the road 3 miles west Winters, and up the Berryessa Road give weight to the pla ment of the epicenter west of both towns. Since Putah Cr flows east–west, landsliding along its banks could have occur both west and east of town, but none was reported to the e In addition, the degree of fissuring decreased rapidly toward east; the report from the Esparto–Capay area east of Vacav and Winters indicates fissures in only "several places". Thus, epicenter of the April 19, 1892, earthquake has been located N of Vacaville and SW of Winters in the ridges bordering Central Valley.

Based on the degree of fissuring and other natural phenom as well as structural damage in Vacaville and Winters, a m imum intensity of VIII–IX (MM) has been assigned to t earthquake.

Duration and Nature of Shaking

As expected, there is no consensus on perceived duration shaking. The estimates ranged from a few seconds to 3 minu (table 2). There is also no great correlation between distan from the epicenter and duration, except that towns m removed from the epicenter reported 30 and 60 seconds for t shaking. In general, it was acknowledged that this earthqua was of notable duration.

> "Professor Davidson ... soid thot for the durotion it was the heavi shock we hove hod since the well–remembered one of 1868. The length it is no doubt responsible for the domoge done in the towns obove m tioned." (*Sonoma Democrat,* April 23, 1892, p. 3)

Table 2

RIL 19, 1892 EARTHQUAKE DIRECTION & DURATION OF SHOCKS (AS REPORTED IN NEWS ITEMS)

Town	Direction of Shocks	Duration of Shaking
eda	NE & SW?	
rado		20 Sec.
och	W→E[a]	40 Sec, 15 Sec.
rn	Northerly, NE & SW, S→N	
cia	N & S[b]	Several mins., several sec.
eley	WNW & ESE	
s	NE & SW	
on City	NW & SE	30 Secs.
o	NW→SE	
sa	N→S then E & W, E & W then N & S	20 Secs.
field	NE & SW	
s Valley	N→S, E→W, SE→NW	Prob. 10-15 Sec.
da	SE→NW	30 Sec.
dsburg	S→N	
	N→S	30 Sec.
use	N→S	Almost 30 Sec.
inez	W→E, N→S, E & W	35 Sec.
sville	NNW & SSE	7-9 Sec., 70-90 Sec.
s Station	N→S	20 Sec.
on	SW→NE	
Hamilton	E & W?	
	E & W, NE→SW	Several Sec.
da City	NE→SW, S→N	1 Min.
astle	W→E	
and		58 Sec.
ge Vale	NW→SE	
nd	N & S	
rmo	N & S	
luma	E→W	3-4 Sec.
erville	E→W	

Town	Direction of Shocks	Duration of Shaking
Reno, Nev.	N→S	
Sacramento	N→S, verging on NW & SE NW→SE, SSW & NNE and Circular[c]	30 Sec.
St. Helena	S→N	20 Sec.
San Francisco	From S	A few Sec.
San Leandro		10 Sec.
San Rafael	N→S	15 Sec.
Santa Cruz	E & W	50 Sec.
Santa Rosa	E→W	Over a minute, fully 3 mins.
Suisun	NW→SE	Fully 1 min.
Sutter Creek	E & W	
Turlock	N→S	A number of seconds
Vacaville	SW→NE then N & S, S→N, SE→NW, SW→NE & NS	3/4 Mins.
Virginia City, Nev.	E→W	
Wheatland	W→E	
White Rock		8 Sec.
Willows	N→S	15 Sec., 1 min.
Winters	E→W	5 - 6 Sec.
Woodland	N & S then EW	
Yuba City	SE→NW	
Light Houses		
East Brother L.H.	E→W	47 Sec.
Fort Point L.H.		7 Sec.
Lime Point L.H.		8 Sec.
Mare Island L.H.	W→E	30 Sec.
Oakland Harbor L.H.	W→E	5 Sec.
Roe Island L.H.		45 Sec.
Yerba Buena L.H.		58 Sec.

W→E indicates shaking with reported direction of origin from the west.
N & S indicates shaking in the north-south direction.
NW→SE, SSW & NNE and circular indicates conflicting reports of the origin of shaking.

	NW-SE	NE-SW	N-S	E-W
Direction at towns reporting 1 direction	6	4	15	12
Direction at towns reporting conflicting directions	2	6	9	5
Total	8	10	24	17

Approximately half of the towns and lighthouses included in their descriptions of the event a judgement as to direction from which the shock waves hit as well as perceived orientation of shaking (see table 2). The reported direction of shaking was almost evenly divided between E–W and N–S, although slightly more locations reported a N–S orientation. Many towns stated that during the event the shaking switched direction to 90° of the initial shock. Some of the news items also described the motion as a distinct circular twisting.

Magnitude

Utilizing the determined epicentral intensity of VIII–IX (MM) and the Gutenberg–Richter (1956) relationship ($M = 1 + \frac{2}{3} I$), a magnitude of 6.7 has been estimated for the April 19, 1872 earthquake.

APRIL 21, 1892, EARTHQUAKE

Intensity Data, Isoseismal Map, and Felt Area

Eighty–two towns and four lighthouses reported this earthquake to the media recovered. In contrast to the reports from the April 19 event, only 16 of these locations had not been included in the Perrine and Holden catalogues, and more of the reports were vague to such a degree as to make intensity designation difficult. The reports tended to concentrate on effects within the area of maximum intensity and to present vignettes and other special features. Table 3 lists the intensities assigned by this author to effects from the April 21 event for each reporting location.

The isoseismal map derived from this information is presented in figure 2. The felt area for the April 21 event extended from at least Half Moon Bay on the west to Virginia City, Nevada, on the east and from Centerville on the south to Redding in the north. Total felt area estimated from these reporting limits is similar to that of the April 19 event, slightly over 90,000 sq. miles.

The isoseismal pattern produced during this event generally resembled that of the April 19 earthquake. The area of highest intensities, IX, VIII, VII, and VI (MM) was oriented nearly N–S, while the areas of Intensity V and IV had a NW–SE elongation, being controlled primarily by the intensities in the Fresno and western Nevada areas. However, in this earthquake, the high intensity area had shifted northward from that of April 19, and the area of intensity VI did not extend as far eastward, although intensities of up to V were felt further to the south, at least to Centerville. In general, the event of April 21, 1892, produced a more N–S elongated isoseismal pattern than did that of April 19.

Epicenter and Maximum Intensity

Reports of the April 21 event from the Winters–Vacaville area indicate greater additional structural damage in Winters than in Vacaville:

"Winters suffered mast severely fram yesterday's shack, thaugh neighboring tawns did nat by any means escape." (*Son Francisco Examiner,* April 22, 1892, p. 1)

This may have been partially due to the fact that Vacaville had already experienced the most substantial wreckage two days earlier. Nevertheless, general comparisons of the two major earthquakes at each town also differed. The report from Vacaville stated,

"Anather earthquake ... nat sa severe as thase af preceding days." (*Record Union,* April 22, 1892, p. 1)

While from Winters,

"Yesterday's shock at Winters was much mare vialent than its wa predecessar." (*Socramento Evening Bee,* April 22, 1892, p. 1)

These observations were explained as follows:

"It is quite clear that the focus af [the April 21] disturbance has bee shifted ta the narth, and has been located near Winters. Elmira an Vacaville gat aff lightly." (*Morning Call,* April 22, 1892, p. 8)

However, particularly because of the already existing al and number of wrecked or weakened structures, the judgm quoted above are by themselves questionable in value. Exami tion of the numerous eyewitness accounts of natural phenom produced by the earthquake yields ample data from which location of the epicenter and the maximum intensity have b determined.

"It must have been a fearful shock. The sand bars in Putah Creek [t the west and sauth af Winters] apened, and fram the fissures the wate spurted high up an the banks. In same places the creek became dry, i athers it changed ta a tarrent. The banks caved in in same places an almast dammed the stream.

"On Mr. Lune's place, a few miles fram Winters, the sand came up fra the battam af the well and filled it. On an adjaining farm, a well was le perfectly dry. In many the water became thick and an unpleasant adar a sulphur rase fram it. The water was sa strangly charged with minerals a ta undrinkable far a while. This, hawever, soan possed away." (*Sa Francisco Examiner,* April 22, 1892, p. 1) af the earth, which rase in billaw aff taward the hills. Then came the shock like a terrible twist, that tare th strangest walls, and dashed dawn enarmaus masses af rack fram th hillside ... Men driving alang the cauntry roads were afraid ta continu their jaurney, far the graund rase in ridges befare the harses and fright ened the beasts until they refused ta proceed a step further." (*Socarmento Evening Bee,* April 22, 1892, p. 1)

"An eyewitness furnished a brief but vivid descriptian af the passag af this wave an Thursday marning. He said that he was cultivating a fiel a little ta the west af the Devilbiss residence, when he naticed that th cultivatar, an which he was riding, plunged vialently. At the same mamen there was a laud, roaring naise, and a claud af dust sweeping rapidly alang taward the tawn af Winters. The graund rase and fell like the sea in a starm, and a mament later a tremedaus crash annaunced that it has struck the Devilbiss hatse. Successive crashes shawed when it reached ather hauses as it possed alang, and when it reached the tawn the naise was tremendaus. Immediately after the shock the eyewitnesses saw distinctly that the graund was braken up inta reefs and furraws, which clased up even while he was watching them. Large valumes af water spurted up fram the dry bed af Putah Creek.

"The greatest damage dane by the earthquake autside af that wraught in the tawn itself is in the wrecking af the cauntry road skirting Putah Creek, an the sauth side af the tawn. A new raad will have ta be cut thraugh adjaining praperty, as a dangeraus cliff fifty feet deep has been farmed at the spot where the embankment gave way. A lang zigzag fissure six inches wide extends right alang the center af the raadway. As in nearly every ather case the fissure runs narth and sauth." (*Morning Call,* April 22, 1892, p. 2)

"Many peaple drave aver [ta Winters] fram Woodland and Dixan. All these say that alang the roads are evidence af the earthquake's power. On the Salana and Pleasant Valley roads particularly the highway is cavered by great cracks and the earth is banked up in hummocks. Near the Walfskill place, where a fine residence was badly shattered, a bluff had fallen and filled up the road. It was reported that several boiling springs had burst fram the foathills an the narth and west and were flawing steadily." (*San Francisco Examiner,* April 22, 1892, p. 1)

And as a summary of the above:

"Fram the general appearance af things .. there seems ta be little daubt that the shock ariginated in the foathills an the west wide ..." (*Morning Call,* April 23, 1892, p. 2)

Based on these reports, the epicenter of the April 21, 1892 earthquake has been placed north of the April 19 epicenter, t the west of Winters in the hilly ridge area, within the ar

Figure II

ISOSEISMALS OF APRIL 21, 1892 EARTHQUAKE

Table 3

APRIL 21, 1892 EARTHQUAKE INTENSITIES

Town	Intensity
Acampo	V-VI?[b]
Alameda[a]	V
Antioch	V?
Auburn	IV?
Benicia	V
Berkeley	
Biggs	VI
Blacks	VII
Camptonville[a]	IV?
Capay[a]	VII?
Carson, Nev.	IV?
Centerville	V
Chico	V-VI
Colusa	VI
Cosumnes[a]	V?
Courtland[a]	V-VI?
Davisville	VII
Dixon	VII?
Downieville	V
Dunnigan	VI
Elk Grove[a]	V
Elmira	VI
Esparto	VIII
Fairfield	VII
Floren[a]	V?
Folsom[a]	IV-V
Fresno	V?
Galt[a]	V-VI?
Gearytown	V?
Georgetown[a]	?
Gold Run	V?
Grass Valley	VI-VII
Gridley[a]	V
Hayward	V
Healdsburg	VI
Knights Landing[a]	VI
Lathrop	IV-V?
Livermore	?
Lodi	VI
Madison	VII
Martinez	VI
Marysville	V-VI
Maxwell	VI?
Merced	?

Town	Intensity
Mills College	?
Mount Hamilton	?
Napa	VI-VII
Nevada City	IV-V
Newcastle	IV or V?[c]
Nicolaus	V?
Oakland	V
Orland	IV?
Oroville	V?
Petaluma	VI
Pinole[a]	V
Placerville	V?
Plainsfield[a]	?
Red Bluff	V
Redding	III-IV?
Reno, Nev.	IV?
Sacramento	VI-VII
St. Helena	V-VI
San Francisco	V
San Jose	III-IV
San Leandro[a]	V
San Rafael	V
Santa Rosa	VI-VII
Sobrante[a]	V
Sonoma	V-VI
Spanishtown	III-IV?
Stockton	V
Suisun	VI
Truckee	IV or V
Vacaville	VII
Virginia City, Nev.	IV
Walnut Grove	?
Williams	VI-VII
Willows	VI-VII
Winters	IX
Woodland	VII
Yolo	VII
Yuba City	V
Light Houses	
Angel Island L.H.	?
Bonita Point L.H.	IV?
Lime Point L.H.	?
Roe Island L.H.	?

a. Towns not reported by Perrine & Holden
b. V-VI indicates an intensity greater than V and less than VI
c. IV or V indicates conflicting reports of intensity

ed on the isoseismal map as intensity IX (MM). Since s from towns in the hill area west of Winters are not ble, it is uncertain how far west the epicenter actually was.

naximum intensity of IX (MM) has been assigned to this compared with VIII–IX (MM) for the earlier event be- of the large–scale hummocky waves in the ground and lly more violent descriptions of all the natural phenomena s landslides, rockslides, ejection of water, and fissuring.

Duration and Nature of Shaking

le 4 lists the perceived duration of shaking of the April 21 as given in the reports. These estimates range from a few ds to about a minute, although most reports indicate time of 30 seconds or less. There were no reports from the ter nor from the most remote towns, except Virginia City. pected, there was little agreement of duration even among boring towns. However, the reports in general agreed:

shock was considered as violent as the first ones, but of less n in time." (*Grass Valley Union,* April 23, 1892, p. 3)

rty–seven locations reported direction of shaking (See ta- . The perceived orientations were evenly divided between and N–S as was the case with the April 19 events. A report arizing the effects within the epicentral area makes an vation which may indicate the orientation of principle sur- eformation:

in nearly every other case, the fissure runs north and south." *ng Call,* April 23, 1892, p. 2)

April 21, 1892, earthquake differed from that of the 19th t there was virtually no conflict among reports from each ular location of orientation of shaking, and there were no s of the twisting motion or switch in shaking direction that perienced during the previous major event.

Magnitude

magnitude of the April 21, 1892, earthquake is estimated ve been 7.0 based on value of IX (MM) for the maximum ity.

APRIL 29, 1892, EARTHQUAKE

s was the last earthquake reported for the late April–early 1892 period in central California. Since this event was ed in the same general region as those of the previous 10 and appears to have been larger than any shock since April investigation relative to the events of April 19 and 21 was taken.

neteen towns reported this event to the newspapers re- d. Most locations described the shock only in terms such as t", "sharp", or "very distinct". Since these adjectives had utilized to described effects during the earlier events of ities from III to VI (MM), they are considered too impre- r use in intensity determination. Thus, few reporting loca- have been assigned intensities for this event, although a al idea of felt effects may be ascertained (see table 5).

m reports of the April 29 event (available), this earth- was assigned a maximum intensity of V–VI (MM) with towns in the epicentral area probably experiencing no more intensity V (MM). The center of the disturbance appears ve been in the same general vicinity as those of the previous s, perhaps a bit south and west of them, judging from the vely strong shaking experienced in this event at Fairfield and Benicia. The felt area does not extend anywhere as far north or south as the previous shocks, but generally has a large east–west extent. The farthest places from the epicenter reporting the earthquake were Stockton to the south, Grass Valley and Marysville to the north, Carson City to the east, and San Francisco and San Rafael to the west. (See figure 3)

Since the felt effects in few towns near the epicenter are recorded, and in general reports are vague, an isoseismal pattern is difficult to determine. It is apparent, though, that the towns reporting highest intensities are oriented in a generally NE–SW line from Benicia through the Vacaville–Winters area to Davis.

The duration of this event was much less than those of the previous larger events; reports ranged from 1½ to 20 seconds. Only seven towns reported perceived direction of shaking, with no consensus among reports.

DISCUSSION AND CONSLUSIONS

Through the investigations outlined above, it has been determined that the epicentral region for the major events of the April 19–29, 1892, series of seismic events in central California was not in the Central Valley, as has been previously assumed, but rather in the hills bordering the valley west of Vacaville and Winters. Such definition precludes the likelihood that any structure within the Central Valley was responsible for these events. Further geologic mapping and geophysical work will be required to provide positive correlation of this series of earthquake with responsible structure.

Analysis of contemporary reports from the epicentral region confirms the maximum intensity of IX (MM) for the April 21 event, as previously listed in "Earthquake History of the United States". It was found, however, that the April 19 event was slightly less severe, with maximum intensity of VIII–IX (MM). In contrast, the felt area of the April 19 earthquake was slightly greater than that of April 21, although both earthquakes were felt from the Pacific coast, south to at least Fresno and throughout west–central Nevada. The isoseismal configuration of both events had a marked NNW–SSE orientation, and in general the events were similar, suggesting a common origin.

The definition provided the April 19–29, 1892, earthquakes by the information presented in this paper and future field investigations, while producing a more realistic picture of California's seismic history, is also expected to render these events more useful than previously in seismic safety planning and engineering design in west–central California.

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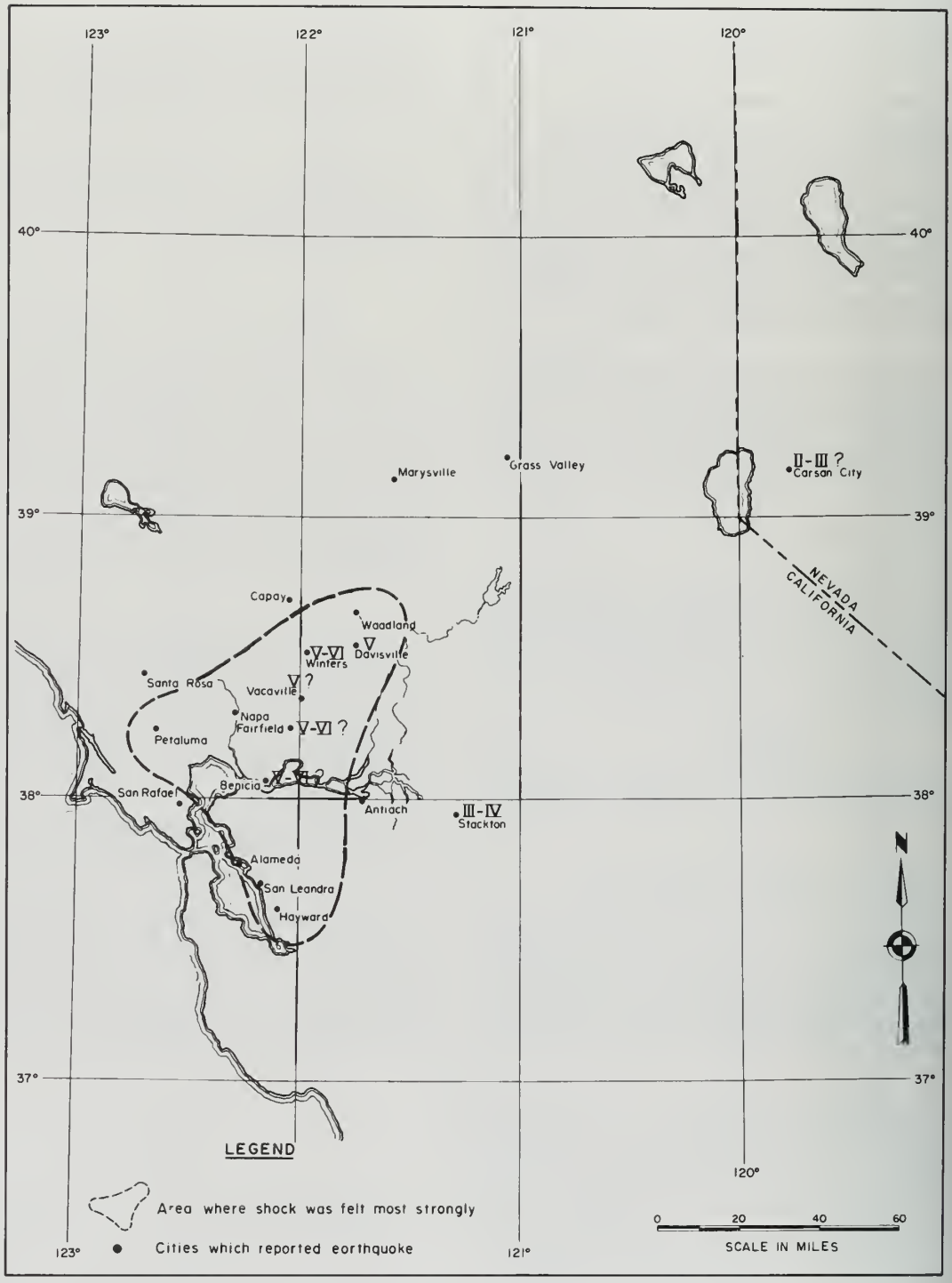

Figure III

APRIL 29, 1892 EARTHQUAKE

Table 4

RIL 21, 1892 EARTHQUAKE DIRECTION & DURATION OF SHOCKS (AS REPORTED IN NEWSPAPERS)

Town	Direction of Shocks	Duration of Shocks
po	E→W[a]	30 Sec.
och	N→S	60 Sec.
eley	N & S[b]	10 Sec.
s	N & S	8 Sec.
erville	N→S	
o	N→S	30 Sec.
sa	E→W	At least 10 Sec.
sville	SE→NW	No more than 5 sec.
no	NW→SE	
s Valley	E→W	Several sec.
ley	N→S	20 Sec.
	NE→SW	15 Sec.
inez	NW→SE	35 Sec.
sville	N & S	
ell	E→W	30 Sec.
s College	E & W	
t Hamilton	E & W	
	E & W, E→W	3/4 Min.
da City	S→N	45 Sec.
laus	NE & SW	20 Sec.
and	E→W	Not quite 30 sec., 20 sec.
nd	NE & SW	8 Sec.

Town	Direction of Shocks	Duration of Shocks
Oroville	N→S	12 Sec.
Petaluma	E→W	Quick
Placerville	W→E	
Reno, Nev.	NW→SE	
San Francisco	E→W, N & S, NE→SW[c]	Short, 13 sec., 20 sec., 3-6 sec.
San Jose		Only a couple of seconds
San Rafael	N→S	
Santa Rosa	E→W	15 Sec.
Sonoma		30 Sec.
Stockton	N→S	10 Sec.
Vacaville	E & W	
Virginia City, Nev.	NE & SW	Short
Walnut Grove	N & S	30 Sec.
Willows	N→S	
Woodland	from SE, E→W	30 Sec.
Yolo	from SE	
Light Houses		
Lime Point L.H.		2 Sec.
Roe Island L.H.	NW→SE	1 Min. 10 Sec.

E→W indicates shaking with reported direction of origin from the east
N & S indicates shaking in the north-south direction
E→W, N & S, NE→SW indicates conflicting reports on the origin of shaking.

	NW-SE	NE-SW	N-S	E-W
Direction at towns reporting on direction	5	4	11	12
Direction at towns reporting conflicting directions	1	1	3	2

Table 5

APRIL 29, 1892 EARTHQUAKE

Town	Intensity	Direction	Duration
Antioch[a]	Light		
Benicia	V-VI? Very severe - no damage		
Capay[a]	?		
Carson City	II-III?	SE & NW	Very short
Davisville	V Slight, loose bricks down		
Fairfield	V-VI? Heavy - no damage		
Grass Valley	No damage		10 Sec.
Hayward	Sharp	E-W	
Marysville	Slight		
Napa	Sharp - no damage	N-S	Short
Petaluma	Qt. sharp	E-W	
San Francisco	Slight - no damage		
San Leandro	Sharp	"Twister"	1-1/2 Sec.
Santa Rosa	Very distinct	E-W	8 Sec.
San Rafael	Slight		
Stockton	III-IV Light, just enough to move chandeliers		20 Sec.
Vacaville	V? Sharp		
Winters	V-VI? No damage - people alarmed		5 Sec.
Woodland	No damage	N & S	

(a) Towns not reported by Perrine & Holden.

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HRUST PLATES OF THE NORTH–CENTRAL KLAMATH MOUNTAINS EAR HAPPY CAMP, CALIFORNIA[1]

CHRISTOPHER W. KLEIN[2]

ABSTRACT

The Klamath Mountains near Happy Camp are formed of at least three stinct lithologic and structural units which are interpreted as imbricated rust plates. Two of these thrust plates are apparent tectonic subdivisions the western Paleozoic and Triassic belt. The third, and lowermost, unit mprises metasedimentary and overlying metavolcanic rocks of the Galice rmation of the western Jurassic belt. The Galice rocks display an inverted tamorphic gradient and locally are also increasingly deformed moving wards in structure. The formation is overlain along a major regional thrust ult by amphibolite, quartzite and associated ultramafic rocks which have en tentatively regarded as part of the western Paleozoic and Triassic belt. e amphibolite unit in turn is overlain along an apparent thrust fault by a eenschist facies assemblage of argillite, andesitic wacke and arenite, eenstone, marble, and chert which is lithologically equivalent to low–grade rts of the western Paleozoic and Triassic belt. The setting and characteris- s of the Galice Formation indicate that it probably correlates with the hists of Candrey Mountain, which occupy a window in the western Paleo- ic and Triassic belt east of Happy Camp.

INTRODUCTION

Geologic mapping of rocks of the western Paleozoic and riassic belt and the western Jurassic belt near Happy Camp in e north–central Klamath Mountains of California (figure 1) rovides insight into the tectonic patterns of the surrounding gion (figure 2). In this region, it can be shown that the western aleozoic and Triassic belt can be subdivided into thrust units f distinct composition and grade of metamorphism and that the hist units of Condrey Mountain which occupy a window in the estern Paleozoic and Triassic belt are probably the metamorph- equivalent of the Galice Formation (western Jurassic belt).

TERRANES OF THE WESTERN PALEOZOIC AND TRIASSIC BELT

Large areas of the western Paleozoic and Triassic belt in rthernmost California are metamorphic rocks of the almandine–amphibolite facies (figures 1, 2, unit "am"). Typical of ese are biotite–quartz schist (some garnetiferous), quartzite, nticular bodies of marble, and fine–to–medium grained amibolite, the igneous antecedents of which apparently include ws, pyroclastic deposits, and tuffaceous sediment. Ultramafic cks are abundant in the amphibolite terranes and are principal- harzburgite and dunite, which ranges from scarcely to comtely serpentinized. Some form large sills or sill–like bodies nich were emplaced in a hot but crystalline state during reonal metamorphism (Medaris, 1966; Barrows, 1969). These ade into smaller lenses and belts (for example in the north of ure 1) which are possibly remnants of larger bodies which ve been infold with and faulted into the rocks which they trude.

Some of the foliated amphibolite masses near Happy Camp ave been intruded by tonalite which has also been metamorhosed. The tonalite forms dikes, sill–like bodies, and tectonic lenses which show an intense syn–metamorphic deformation. Locally, still younger, nearly undeformed amphibolite dikes cut the meta–tonalite, older amphibolite, and metasedimentary rocks. Rocks mapped as part of the amphibolite unit at Happy Camp also include some of similar composition which are metamorphosed only to the greenschist facies (figure 1, areas within "am" in the extreme northwest and east of Happy Camp). The greenschist and amphibolite facies rocks are apparently gradational.

The age of amphibolite at Happy Camp is unknown, but it appears likely that the rocks correlate with amphibolite east of Condrey Mountain (figure 2), which underwent Jurassic metamorphism (K–Ar ages of hornblende of 148 and 146 m.y., Lanphere and others, 1968).

The amphibolite of the western Paleozoic and Triassic belt in the north–central Klamaths has been described as gradational with rocks of lower metamorphic grade in adjacent regions of the belt (e.g., Hotz, 1967; 1971, p. 12). However, amphibolite at Happy Camp is also faulted beneath a greenschist metamorphic facies assemblage of argillite, abundant andesitic arenite (or wacke), marble, greenstone, and chert (figure 1, unit "m"), which is of unsuitable composition to be a parent of rocks of the amphibolitic terrane. Serpentinite exposed in an east–trending zone across part of the greenschist facies unit may occupy a fault zone which bisects it. However, ultramafic rocks are absent from the unit elsewhere.

The fault between the greenschist facies unit ("m") and underlying amphibolite ("am") is interpreted to be an east–dipping thrust parallel to structures in both units. Rocks of the greenschist facies unit are very similar to rocks of the western Paleozoic and Triassic belt (or Applegate Group) in Oregon immediately north of Condrey Mountain described by Engelhardt (1966), and Heinrich (1966). Structural relationships between those rocks and amphibolite east and west of Condrey Mountain (figure 2) have not been investigated. However, to the east and south of Condrey Mountain, Hotz (1967) recognized in the western Paleozoic and Triassic belt fault–bounded terranes of distinct composition and metamorphic grade. Thus it appears that the distinct units of the western Paleozoic and Triassic belt at Happy Camp may be two (or more) thrust plates which form the belt in the north–central Klamath Mountains.

GALICE FORMATION

The Galice Formation which underlies the central Happy Camp area (figure 1) occurs in a north–trending anticlinorium which has a well–developed asymmetric metamorphic gradient. Rocks in the west are greenstone overlying interbedded slate and metasiltstone, fine–grained schistose metagraywacke (textural zone 2 + and 3–), minor foliated pebbly wacke and arenite; to the east, the metasedimentary rocks grade upward into phyllite and fine–grained quartz–mica schist, which are overlain and interbedded with greenstone and greenschist which in turn grade upward (and eastward) into epidote amphibolite. These mafic rocks have been previously mapped to the east and northeast

eviewed by Preston Hotz and Porter Irwin.
eothermEx, Berkeley, California.

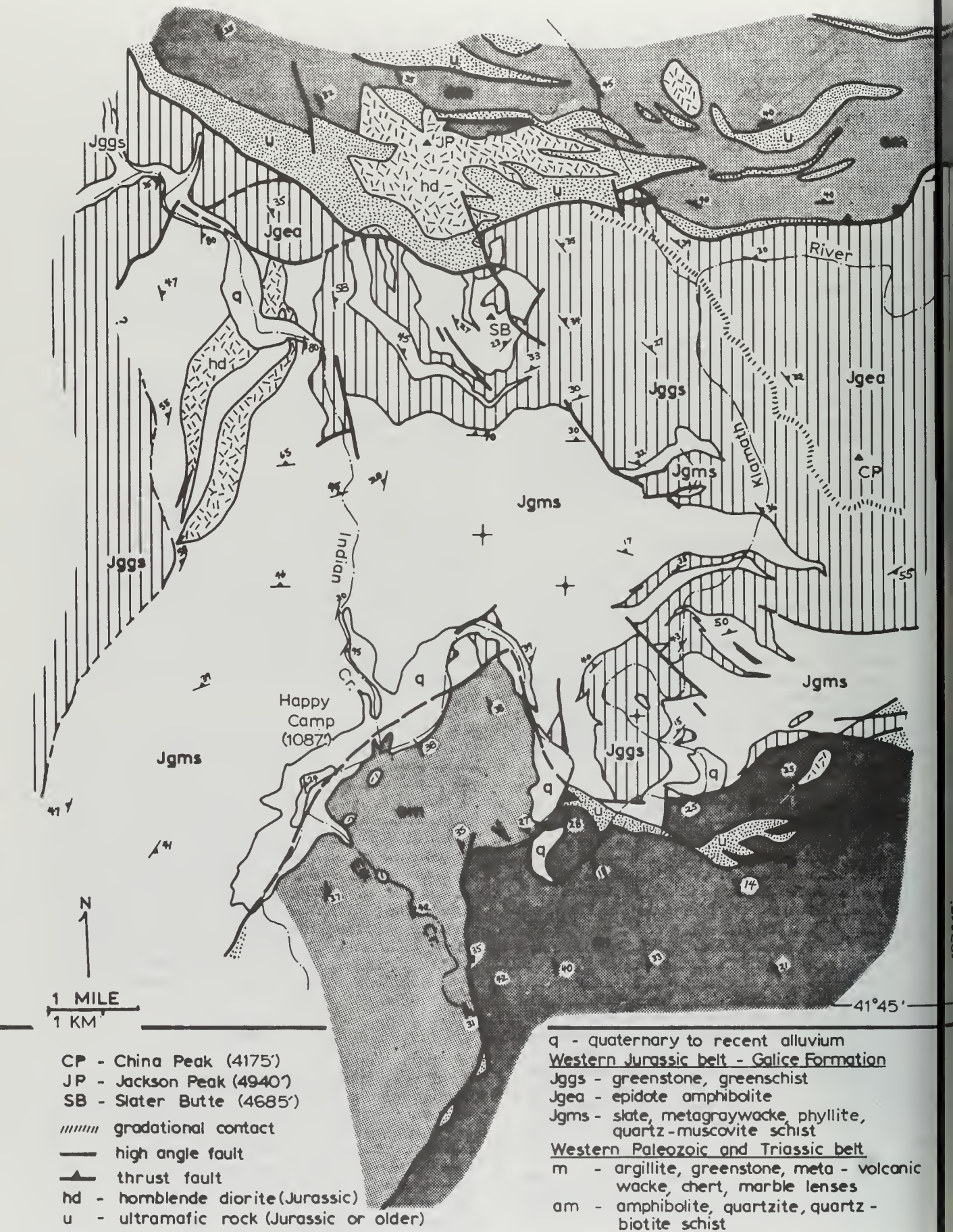
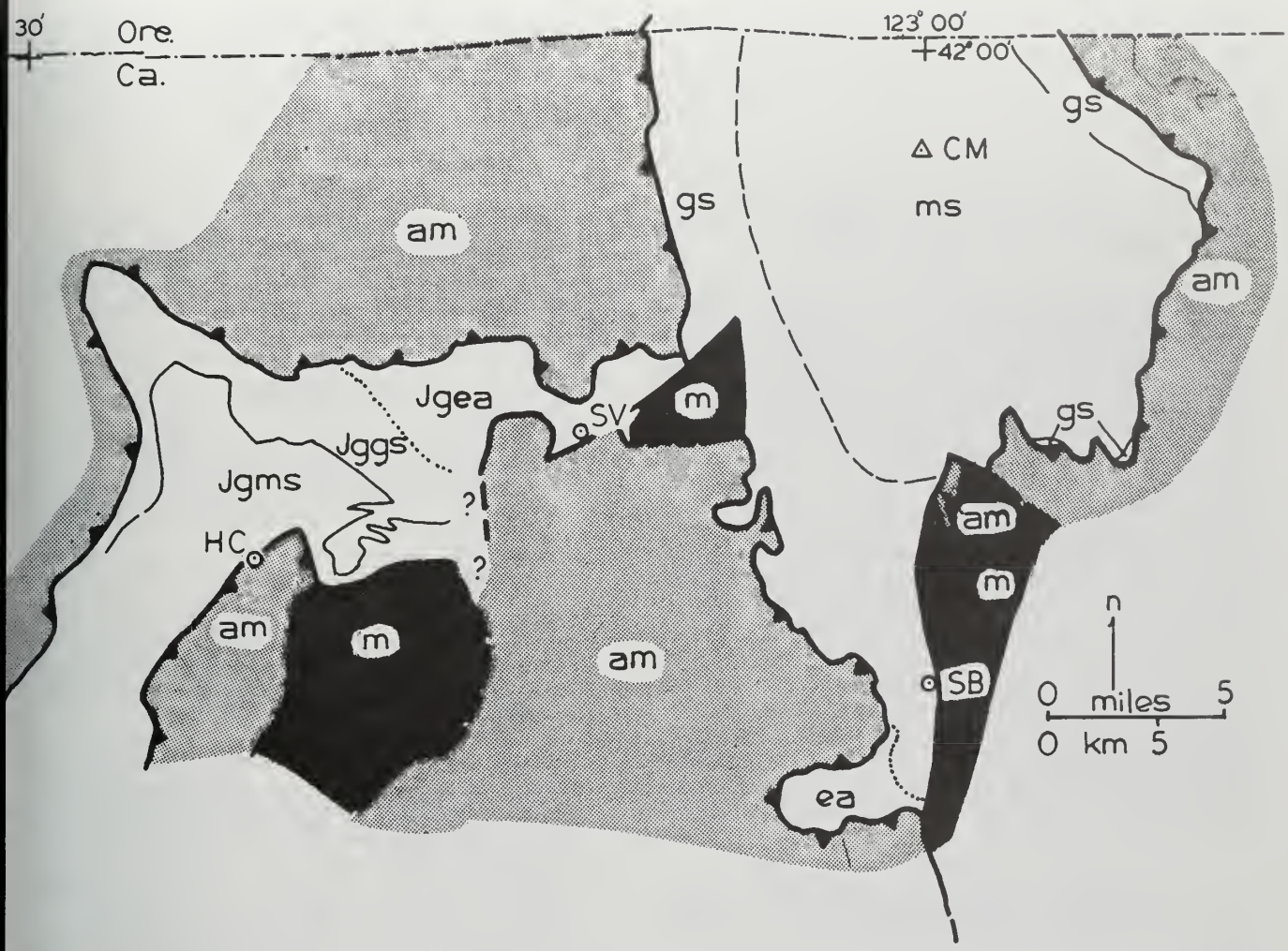


Figure 1. Geologic map of the Happy Camp area.



gure 2. Generalized geologic map of the north–central Klamath Mountains. Heavy toothed lines: contacts between thrust plates. Light lines: ratigraphic contacts. Dotted lines: metamorphic isograds. CM: Condrey Mountain. HC: Happy Camp. SB: Scott Bar. SV: Seiad Valley. am: mphibolite, quartizite, marble, ultramafic rocks. m: argillite, volcanic sediments, greenstone, marble, chert of the western Paleozoic and Triassic lt. Jgms, Jggs, Jgea: Galice Formation metasediments, greenschist, epidote amphibolites. ms, gs, ea: mica schists, greenschists, epidote mphibolites of Contrey Mountain. Jurassic granitic intrusives not shown.

ear Seiad Valley (Dickinson, 1959; Medaris, 1966; Rynearson nd Smith, 1940); however, they have not previously been recog-ized as part of the Galice Formation. In the Seiad area, Medaris 1966) also noted that greenschist grades upward into epidote mphibolite. In parts of the metasedimentary terrane (for exam-le, on Slater Butte), the inverted metamorphic gradation is ccompanied by post–metamorphic fold–deformation that also ncreases upward in structure.

STRUCTURE

The Galice Formation at Happy Camp underlies the am-hibolite unit along a thrust fault which has been modified by eeply dipping faults. The original thrust fault appears to be reserved only in the northeast of figure 1, dipping northward arallel to metamorphic foliation in the units above and below. Within figure 1 and to the east, the Stanford Geological Survey Dickinson, 1959) and Medaris (1966) interpreted this contact s an isograd between epidote amphibolite to the south and mphibolite to the north. The discovery that the epidote am-hibolite is part of the Galice (whereas the amphibolite belongs the western Paleozoic and Triassic belt) implies instead that he contact is a fault, an interpretation which Medaris (pers. omm., 1973) also now favors.

ORIGIN OF SCHIST AT CONDREY MOUNTAIN

The origin of low–grade schist which underlies a region around Condrey Mountain, east of Happy Camp (figure 2, units "ms, gs, ea"), has long been uncertain. The schist of Condrey Mountain consists of a core of quartz–muscovite schist overlain and surrounded by an outer discontinuous ring of actinolite–chlorite schist. The schist apparently occupies a window in the western Paleozoic and Triassic belt, as it is overlain by amphibolite and ultramafic rocks of the belt along a ring–shaped system of moderately to steeply dipping reverse faults (Barrows, 1969; Hotz, 1967, 1971; Kays, 1968; Medaris, 1966; Strand, 1963).

The Galice Formation has a suitable composition and structural position to make it a plausible parent of the Condrey schist; however, its apparent age poses a problem. Lanphere and others (1968) obtained a 141 m.y. age by potassium–argon determination on mica in schist from the Condrey terrane; and Suppe and Armstrong (1972), who suggested that the schist is a Franciscan equivalent, obtained from a whole–rock sample a potassium–argon age determination of 155 ±3 m.y. These ages are more or less the same as, and notably older than, the age of known fossil localities in the Galice Formation, which are late Oxfordian to

early Kimmeridgian (Imlay, 1961, p. D–10), or about 140 m.y. old (Lanphere and others, 1968, figure 7). But fossils are not abundant in the Galice Formation (there are none near Happy Camp); and, as its structure and stratigraphy are not well understood, it may encompass a greater span in age than currently appreciated. In southwest Oregon, for example, part of the Galice has been intruded by diorite which has yielded K–Ar ages ranging from 137 ± 4 m.y. to 145 ± 4 m.y. (Dott, 1965; Coleman, 1971), which indicates a lower limit of about 145 m.y. for the depositional age of the Galice in that area.

In spite of these possibly conflicting ages, two points strongly suggest that the Galice Formation may be regarded as a suitable parent of the Condrey schist. First, the relationship between Galice pelite and (overlying) greenschist and epidote amphibolite at Happy Camp is equivalent to the association between the pelitic and mafic schists of Condrey Mountain. East of Happy Camp and northeast of Seiad Valley (figure 2), the Galice epidote amphibolite abuts the mafic schist of Condrey Mountain along a high–angle fault (Medaris, 1966). Medaris considered that rocks on opposite sides of the fault belonged to two different map units. However, if the Galice and Condrey Mountain rocks are indeed correlative, this fault lies within a single stratigraphic and tectonic unit and only separates portions of the unit which have undergone slightly different grades of metamorphism.

Second, southwest of Scott Bar (figure 2) greenschist of the Condrey terrane grades upward into epidote amphibolite which is in turn overlain along an apparent thrust fault by amphibolite (Barrows, 1969). This structural sequence and metamorphic gradation upwards within the lower plate is equivalent to that in the Happy Camp and Seiad Valley areas (this study; Medaris, 1966), and the rocks of both areas are quite similar in character.

If the correlation here proposed is correct, the Galice Formation has undergone a minimum relative east–west displacement beneath overlying amphibolite of about 22 miles at Happy Camp.

Inverted metamorphic gradients in rocks underlying thrust faults are known in the Klamath Mountains and the Franciscan Formation of the California Coast Ranges (e.g., Irwin and others, 1974; Blake and others, 1967) as well as elsewhere (Armstrong and Dick, 1974). Where the metamorphism is of a high–pressure–low–temperature variety, as in the Franciscan Formation, the rocks appear to have experienced a rapid displacement deep beneath a cool upper plate, although "tectonic overpressure" has also been suggested to produce the high pressures necessary for the observed metamorphsim. In contrast, metamorphism of the Galice–Condrey unit and overlying amphibolite represents moderate pressures and temperatures. The inverted gradient in the lower unit may have formed as it was displaced beneath the hotter amphibolites during regional metamorphism.

The amphibolite unit underwent intense syn–metamorphic deformation with simultaneous tectonic emplacement of ultramafic rock and intrusions of tonalite. The presence of nearly undeformed but metamorphosed mafic (amphibolite) dikes the unit shows that its temperature was still elevated follow the internal deformation and older intrusions, and it could h been still quite hot when subsequently the Galice–Condrey u was displaced (relatively) beneath it.

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IMITATIONS ON DISPLACEMENT AND SOUTHEASTWARD XTENT OF THE DEATH VALLEY FAULT ZONE, CALIFORNIA[1]

y GREGORY A. DAVIS [2]

ABSTRACT

e Death Valley fault zane, sauthern Death Valley area, is characterized Quaternary, including Halacene, right-slip displacement. The zane has en the subject af much speculatian regarding its tatal strike-slip displaceent and its extent inta regians sautheast of Death Valley. Field studies in e sauthern Death Valley area indicate that maximum right-slip acrass the ult zane is appraximately 8 km, nat the 50 ta 80 km speculated by same orkers. Despite hypatheses that the zane extends 250 km to the Colorado ver barder btween Arizana and Califarnia, it cannat be traced sauth af the rtheastern Avawatz Mauntains in the sauthern Death Valley regian. Narthest-striking faults expased ar inferred ta exist in the Sado, Bristol, Old Dad, ranite, Riverside, Big Maria, and Little Maria Mauntains cannat be carrelatd with the Quaternary Death Volley foult zane because af older oge, fferent pasitian with respect ta distinctive gealagic terranes, ar the absence demanstrable strike-slip displacement.

INTRODUCTION

The Death Valley fault zone is an assemblage of northwestriking faults which occupy the relatively narrow floor of Death alley south of 36° N. latitude. The 3 to 6 km-wide zone has been escribed by Noble and Wright (1954), Hill and Troxel (1966), nd Wright and Troxel (1967). Although portions of the zone re concealed beneath Quaternary deposits, evidence for Holoene faulting along some strands of the zone has been reported. eatures indicating right-lateral displacement include: (1) offset a Quaternary cinder cone north of Shoreline Butte (Noble and Wright, 1954); (2) offsets of stream channels south of Mormon oint along the west side of the Black Mountains (Hill and roxel, 1966); (3) drag folds in Tertiary to Holocene sediments Hill and Troxel, 1966); and (4) an offset fold axis in Tertiary rata (Wright and Troxel, 1967).

Although investigators of the zone accept right-lateral disacement along it, considerable controversy exists concerning 1) the magnitude of total strike-slip displacement (see Wright nd Troxel, 1970, *vs.* Stewart and others, 1970), and (2) the outheastward extent of the fault zone beyond the southern eath Valley area. The controversy stems in large part from tewart's estimate (1967) of 80 km of right-slip displacement cross a combined Death Valley-Furnace Creek fault zone, and amilton and Myers hypothesis (1966) that the Death Valley ault zone extends 250 km southeastward into mountain ranges orth of Blythe, California, along the Colorado River (figure 1). amilton and Myers further stated that the Death Valley zone robably has about 50 km of right-lateral displacement in southrn Death Valley. Excerpts from their discussion of the regional xtent of the Death Valley fault zone follow (1966, p. 530–531):

> Faults af the Death Valley zane are expased in the regian sauth af the arlack intersectian in such ranges as the Avawatz, Soda, and Old Dad . . . . The Death Valley fault zane passes thraugh a regian in which orthwest-trending faults separate cantrasted terranes (Bishap, 1964) ta onnect with faults mapped by Hamiltan (unpublished data) in the Big aria, Little Maria, and Riverside mauntains (alang the sautheastern part f the fault zane within Califarnia), where right-lateral faults are spread hraugh a zane at least 20 km wide. Single faults in this part af sautheastern alifarnia have right-lateral displacements af 2–15 km each and an aggreate displacement prabably af about 50 km."

Subsequent studies of the fault zone in the southern Death Valley area, reviewed briefly below, have generally concluded that total right-slip along it has not exceeded about 8 km. Furthermore, it is now known that the faults referred to by Hamilton and Myers in the Big Maria, Little Maria, and Riverside Mountains are too old to correlate with the active Death Valley fault zone. These faults are unconformably overlain by latest Miocene to Pliocene sediments of the Bouse Formation, a shallow marine to lacustrine unit deposited in the Colorado River trough. Hamilton apparently recognized this important age relationship in his later (1971) conclusion that the northwest-striking faults of the region north of Blythe are of "middle(?) Tertiary" age.

Nevertheless, the twin concepts of large displacement and a continuation of the Death Valley fault zone far to the south of its Death Valley occurrence persist in the literature. Stewart and Poole (1974, figure 2), for example, show a regional Furnace Creek-Death Valley fault zone that extends 550 km from Esmeralda County, Nevada, to the Colorado River near Blythe. The same authors (1975, figure 6) project the Death Valley zone 150 km to the south of Death Valley and show approximately 50 km of right-lateral displacement across it of the southeastern margin of the Cordilleran miogeosyncline. This paper describes briefly the results of field studies undertaken to evaluate the existence of a regional Death Valley fault zone of Quaternary age in southeastern California.

DISPLACEMENT ON THE DEATH VALLEY FAULT ZONE, SOUTHERN DEATH VALLEY

Wright and Troxel (1967) present several lines of evidence that right-lateral displacement along the Death Valley fault zone is no more than about 8 km. Their evidence for cumulative right-slip is based on the limited offset across Death Valley of a number of linear geologic features, among them: (1) the line formed by the intersection of the tilted lower contact of the Precambrian Kingston Peak Formation with the unconformity at the base of the overlying Noonday Dolomite ("nd/Kp", figure 2); and (2) a Precambrian talc-bearing belt about 110 km long which is confined to a carbonate member of the Precambrian Crystal Spring Formation (figure 2).

Additional argument for limited right-slip has been presented by Davis and Burchfiel (1973), who propose that the east-striking Garlock fault zone is offset approximately 8 km along the younger, cross-cutting Death Valley zone. Reconnaissance mapping by the writer of the geologic terrane which lies south of the latitude of the Garlock fault and through which any hypothesized extension of the Death Valley zone must extend (figure 2) has not uncovered any evidence for major strike-slip disruption. In this terrane, which includes the eastern Avawatz and Soda Mountains, the hills north of Silver Lake and the western Halloran Hills, marble of uncertain age ("m", figure 2) is intruded by Mesozoic gabbro, diorite, and quartz diorite ("di", figure 2). This rock assemblage is highly distinctive and geographically confined to the area north of Baker.

Original estimates by Stewart (1967) and Hamilton and Myers (1966) of many tens of kilometers of strike-slip displacement along the Death Valley zone were based on interpretations of isopach data from Precambrian and Cambrian stratigraphic units in the southern Death Valley region. Hamilton and Myers'

Reviewed by John C. Crowell, Mason L. Hill, and Bennie W. Troxel.

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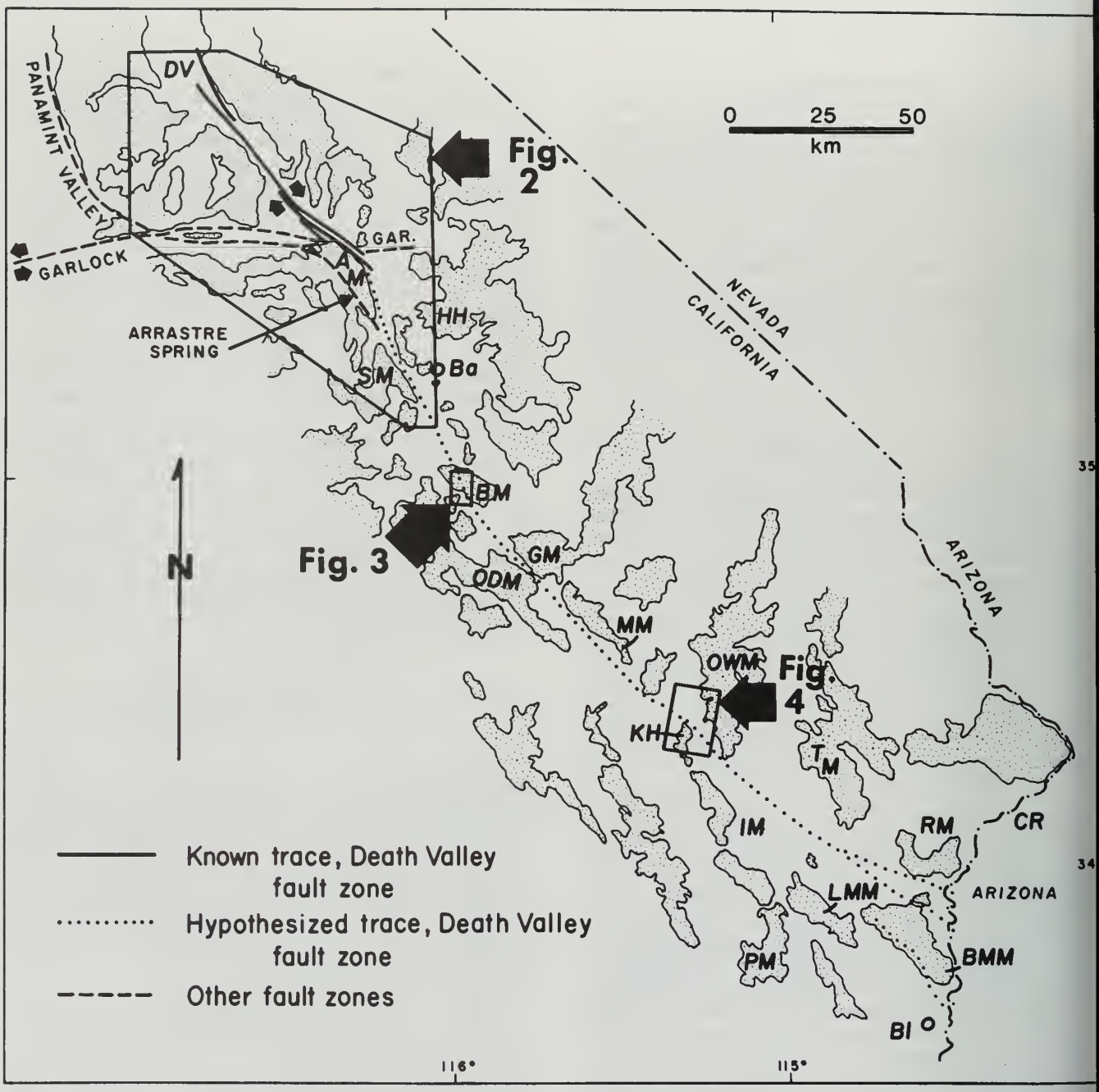

Figure 1. Locotion mop, southeostern Californio, showing known ond hypothesized troces of Deoth Volley foult zone. Geogrophic locolities (north to south DV = Deoth Volley; AM = Avowotz Mtns.; HH = Holloron Hills; Bo = Baker; SM = Sodo Mtns.; BM = Bristol Mtns.; GM = Gronite Mtns.; ODM = Old D Mtns.; MM = Morble Mtns.; OWM = Old Woman Mtns.; KH = Kilbeck Hills; TM = Turtle Mtns.; IM = Iron Mtns.; RM = Riverside Mtns.; CR = Colora River; LMM = Little Morio Mtns.; PM = Polen Mtns.; BMM = Big Morio Mtns.; Bl = Blythe.

timate of 50 km of right-slip on the Death Valley zone, for ample, utilized isopach data of Wright and Troxel (1966). wever, only *one* of the 18 stratigraphic sections referenced by right and Troxel lies *west* of the Death Valley zone. Accordgly, the trend, and therefore the offset, of isopachs west of the ult zone is indeterminate.

Stewart and others (1970, p. 139) in an exchange of ideas with right and Troxel (1970) regarding large lateral displacement the Death Valley fault zone abandoned the earlier, large-slip nclusion of Stewart (1967): "Wright and Troxel (1966, 1967) ve outlined evidence that seems to indicate little slippage on her the Death Valley or Furnace Creek fault zones. If their servations are correct, and we believe they are . . . . " Rent papers by Stewart and Poole (1974, 1975) have, however, mentioned above returned to the twin concepts of large disacement and regional extent for the Death Valley fault zone. evertheless, there is considerable geologic evidence that the ult zone has only minimal displacement (ca. 8 km) in the uthern Death Valley area. The possible extension of this Quarnary fault zone into regions southeast of Death Valley is now nsidered.

POSSIBLE SOUTHEASTERN EXTENSION OF THE DEATH VALLEY FAULT ZONE

If the Death Valley fault zone extends southeastward from eath Valley into the Big Maria, Little Maria, and Riverside ountains as Hamilton and Myers proposed, its position in tervening areas is rather closely constrained by the geographic ttern of basins and ranges and by the presence of known rthwest-striking faults. A regional southeastern extension of e Death Valley zone would presumably include faults in the rtheastern Avawatz Mountains, in the Soda Mountains west Baker, and in the Bristol Mountains farther south (figure 1). or 160 km south of the Bristol Mountains, however, the posated trace of the fault zone lies beneath Quaternary alluvial nd lacustrine deposits that extend without interruption to the iverside, Big Maria, and Little Maria Mountains. If present in is covered interval, the Death Valley zone would have to pass etween closely adjacent bedrock exposures in the Old Dad and ranite Mountains and between the Kilbeck Hills and Old oman Mountains farther south (figure 1). Results of studies each of these geographic areas are now discussed in sequence om north to south.

Soda Mountains

Grose (1959) shows the Death Valley fault zone curving uthward along the eastern edge of the Avawatz Mountains and ining the Soda-Avawatz fault zone in the Soda Mountains west Baker. Although his correlation has not been previously questoned, geologic relations in the Avawatz Mountains suggest that is in error. The distribution of distinctive Precambrian to Mesozoic rock units shown in figure 1 indicates that the Soda-Avawatz fault zone of Grose is a southward continuation of the Arrastre Spring fault zone which lies within and along the southestern border of the Avawatz Mountains. Both of these fault ones (Soda-Avawatz and Arrastre Spring) display a similar attern of branching faults, they are in direct alignment with ach other, and both separate different geologic terranes—unlike he Death Valley zone to the east which lies *within* a distinctive re-Tertiary terrane (figure 2). Late Precambrian to Cambrian eosynclinal strata, primarily quartzitic ("q", figure 2), and narble intruded by mafic plutonic rocks lie east of the Arrastre pring and Soda-Avawatz faults. These units are not exposed west of the faults where metavolcanic rocks of Mesozoic age are videspread.

Grose (1959) describes evidence for right-slip along the Soda-Avawatz fault zone, but regional relations indicate that it and the Arrastre Spring zone to the north must have a considerable, if not predominant, dip-slip component of displacement. Grose states that the contrasting geologic terranes east and west of the Soda-Avawatz zone imply major lateral displacement along it. He concludes (p. 1543) that right-lateral displacement "is probably on the order of 3 miles and possibly much greater", although no evidence for this amount of displacement is offered other than the contrast in fault-juxtaposed terranes. Alternatively, Davis and Burchfiel (1973) regard the western metavolcanic rocks as the probable downdropped Mesozoic cover for the older geosynclinal section which has been uplifted and exposed widely to the east of a Soda-Arrastre Spring fault zone. The Soda-Arrastre Spring fault zone defined here does not comprise a branch of the younger and more easterly Death Valley zone because it, unlike the parallel Death Valley zone, terminates northward at the Garlock fault and may be offset along it some 40 to 50 km in a left-lateral sense (Davis and Burchfiel, 1973). Grose concludes that the Soda-Avawatz fault zone has been active since early Pliocene time, but that its youngest displacements have been postdated by at least two generations of Quaternary sedimentation in the Soda Mountains area.

Differences between the Death Valley and Soda-Arrastre Spring fault zones in age, inferred displacement history, and position with respect to contrasting geologic terranes lead to the conclusion that the two zones are not related and that the Death Valley fault zone does not extend into the Soda Mountains. It seems more likely that the Death Valley fault zone diminishes in lateral displacement southeast of the northeastern corner of the Avawatz Mountains and dies out beneath Quaternary alluvial deposits north of Silver Lake.

Bristol, Old Dad, and Granite Mountains

Faults in the Soda Mountains project southeastward under the Quaternary playa and dune deposits of the Soda Lake basin and toward the Bristol Mountains where several northwest-striking faults are shown on the Kingman (Jennings, 1961) and Needles (Bishop, 1964) sheets of the Geologic Map of California. Not all of these faults are verifiable in the field, and no through-going Quaternary fault of any kind was recognized in the Bristol Mountains. One broad (0.25–0.75 km) northwest-striking zone of sheared and shattered granitic rocks and Tertiary sedimentary and volcanic rocks was mapped in reconnaissance (figure 3). Its sense of displacement is unknown. Although it might conceivably represent a southern extension of the Soda-Arrastre Spring fault zone, it cannot correlate with the younger Death Valley zone because it is overlain unconformably by fanglomerate of late Tertiary or early Quaternary age (Bassett and Kupfer, 1964, plate 1). These alluvial deposits, together with rocks of the underlying shear zone, constitute the footwall of a younger, north to northwest-striking reverse fault which dips eastward beneath a hanging wall of granitic rocks. The fanglomerate strata are locally overturned beneath this reverse fault (figure 3). The fault does not cut or affect either Holocene alluvium or somewhat older, post-fanglomerate alluvial deposits exposed above present stream drainages. There is no basis to correlate the Bristol Mountains reverse fault with active strike-slip faults of Death Valley type.

The Needles geologic map sheet shows a northwest-striking fault beginning in and cutting southward across the fanglomerate unit cited above (see "QP", map sheet designation, Bishop, 1964). This fault is then inferred to extend 35 km to the southeast, concealed beneath Quaternary deposits. The fault cannot be seen on aerial photographs of the southern Bristol Mountains,

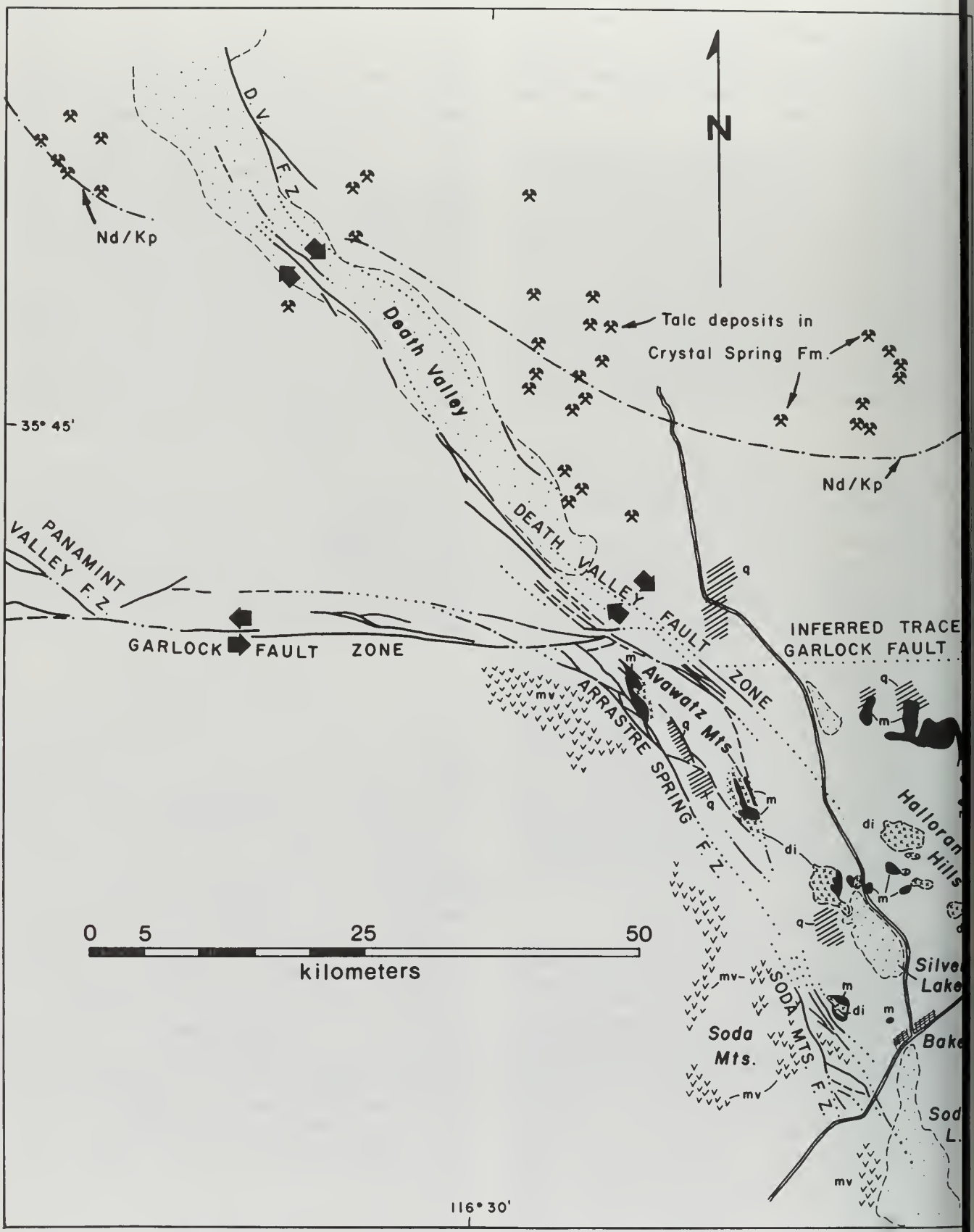

Figure 2. Map of mojar fault zones and selected geologic features, southern Death Valley to Baker area. Rock units: q = late Precambrian-Combrian quartz
m = marble and unmetamorphosed equivolents of uncertain age; di = Mesozoic diorite and gabbro; mv = Mesazoic metavolcanic rocks. Nd/Kp = appr
imate line of intersection, base of Kingstan Peak Fm. with overlying base of Noondoy Dolomite.

r does it offset the unconformity at the base of the tilted and erturned fanglomerate unit.

Ground observations and inspection of aerial photographs of e alluviated pass between the Granite and Old Dad Mountains the south of the Bristol Mountains (figure 1) have yielded no idence for active faulting there, although a fault relationship tween the two ranges seems likely. Granitic and metamorphic cks in the Granite Mountains rise nearly 1000 m higher than enozoic volcanic rocks in the Old Dad Mountains to the west. ocks characteristic of the two ranges crop out within one or o hundred meters of each other in higher portions of the pass, hich is essentially a thinly-alluviated pediment. The northwest-riking contact between granitic and volcanic rocks is presumed be a high-angle fault; it lies along the northeastern base of the ld Dad Mountains. Volcanic rocks closest to this contact have eep variably oriented dips and are locally isoclinally folded. old hinges trend approximately parallel to the inferred fault ntact, but vary in plunge from subhorizontal to vertical within ngle outcrops. The absence of shearing in granitic rocks adja-nt to the inferred fault indicates that the zone of faulting nnot be wider than a few hundred feet and is probably consid-ably narrower. The sense of displacement along the fault in-rred to separate the two ranges is not known, although the ranite Mountains on the east have been elevated with respect the Old Dad Mountains to the west. The age of last displace-ent is also not known, but no faulting of Quaternary units is pparent on aerial photographs of the pass. Several generations alluvial fan deposits at the northern end of the Old Dad ountains have an unbroken northeast-trending grain across e suspected fault as it projects northwestward through the pass rwards the Bristol Mountains.

Kilbeck Hills—Old Woman Mountains

If a northwest-striking Death Valley fault zone (or other ma-r fault) extends southeastward from the pass between the Old ad and Granite Mountains into the Colorado River region igure 1), it must pass beneath alluvial deposits between the ilbeck Hills and Old Woman Mountains (figure 4). Mapping these adjacent and well-exposed areas by W. Evenson and J. nes (1973, unpublished M.S. theses, Univ. Southern Califor-a) indicates the essential continuity between them of rtheast-trending folds, a major low-angle thrust fault, and a stinctive and lithologically varied assemblage of granitic, meta-dimentary, and meta-igneous rocks of Mesozoic, Paleozoic, d Precambrian age (figure 4). Although direct correlation of lds between the Kilbeck Hills and Old Woman Mountains is t possible because of their 5-km separation by alluvium, the acroscopic and mesoscopic structural style of the two areas is rikingly similar. Geologic relations in the Kilbeck Hills and ld Woman Mountains do not preclude the possibility of a rthwest-striking fault beneath the alluvium separating the two eas. However, if a fault does exist, its displacement, whether ip-slip or strike-slip, must be extremely limited. No major fault ith regional implications can pass between the two areas.

Other Considerations

The hypothesized trace of the Death Valley fault zone south the Bristol Mountains coincides with a physiographic linea-ent proposed by Bassett and Kupfer (1964, p. 39). The linea-ent is believed to mark a change in trend of Mojave Desert ountain ranges—ranges to the southwest are described as hav-g northwestern trends, whereas ranges to the northeast trend rtherly or northeasterly. Although generally true, there are ough exceptions to cast doubt on the validity of the proposed neament. The Marble Mountains lie directly east of the linea-ent, but have a pronounced linear, northwestern trend. Despite a northern trend for the entire Turtle Mountains block east of the lineament, these mountains are physiographically and structurally divided into three northwest-trending segments. Ranges with northerly trends southwest of the lineament include the Kilbeck Hills-Iron Mountains and the Palen Mountains. Bassett and Kupfer (1964, p. 39) specifically considered the possibility that their inferred lineament might coincide with a southern extension of the Death Valley fault zone, but they concluded that there was no direct evidence for that zone between the Granite and Old Dad Mountains or farther to the southeast.

Fuis (1976) has recently reported seismological evidence for a major north-northwest-trending tectonic boundary in the eastern Mojave Desert. The boundary, a reasonably well-defined line between seismic (western) and aseismic (eastern) portions of the Mojave region, does not coincide with the inferred trace of a regional Death Valley fault zone nor with any apparent topographic and/or structural lineament in the Mojave Desert. Although the northern end of the seismic/aseismic boundary lies near the Bristol Mountains, its southern end is in the vicinity of the Chocolate Mountains, fully 80 km southwest of Blythe on the Colorado River.

CONCLUSIONS

Hamilton and Myers' suggestion (1966) of a regional correlation between faults in the Riverside, Big Maria, and Little Maria Mountains and the Death Valley fault zone 250 km to the north was a provocative idea appropriate for its time. More recent studies in the Death Valley and Colorado River areas, and in the intervening region between the two presumed end segments of the regional fault zone, necessitate rejection of the Hamilton-Myers hypothesis.

It is concluded here that the Death Valley fault zone does not extend far south of the southern Death Valley area, and that maximum displacement along it is about 8 km, not the 50 to 80 km hypothesized by some workers. Furthermore, the Colorado River area faults with which the active Death Valley zone was to connect are now known to be much too old (pre-latest Micone) to sustain the correlation. Finally, studies in intervening areas along the hypothesized trace of a regional Death Valley fault zone preclude the existence of a through-going strike-slip fault of Quaternary age. No Quaternary deposits along this trace in areas south of the Bristol Mountains exhibit fault disruption, a conclusion supported by the recently published Fault Map of California (Jennings, 1975). This study has not eliminated the possibility of a pre-Quaternary fault zone of regional extent within the area under discussion. Such a zone might include the Soda-Arrastre Spring fault zone and its possible northern counterpart, the Panamint Valley fault zone (Davis and Burchfiel, 1973), the northwest-striking shear zone in the Bristol Mountains, and the fault inferred to separate the Old Dad and Granite Mountains. Connection of this speculative older fault zone to the Colorado River area faults described by Hamilton and Myers is much more problematical. Field relations in the intervening Kilbeck Hills-Old Woman Mountains area strongly refute the passage through that critical region of any major northwest-striking fault zone, whether dip-slip or strike-slip.

ACKNOWLEDGEMENTS

Field studies along the hypothesized southeastern extension of the Death Valley fault zone were undertaken on behalf of Woodward-Clyde Consultants, Orange, California, and the Southern California Edison Company. This work was part of a regional evaluation of tectonic activity in southeastern California with respect to nuclear power plant siting in the eastern desert region. Appreciation is extended to Stephen T. Freeman, Shingi Kuniyoshi, Terry J. Shackelford, and Lawrence M. West of

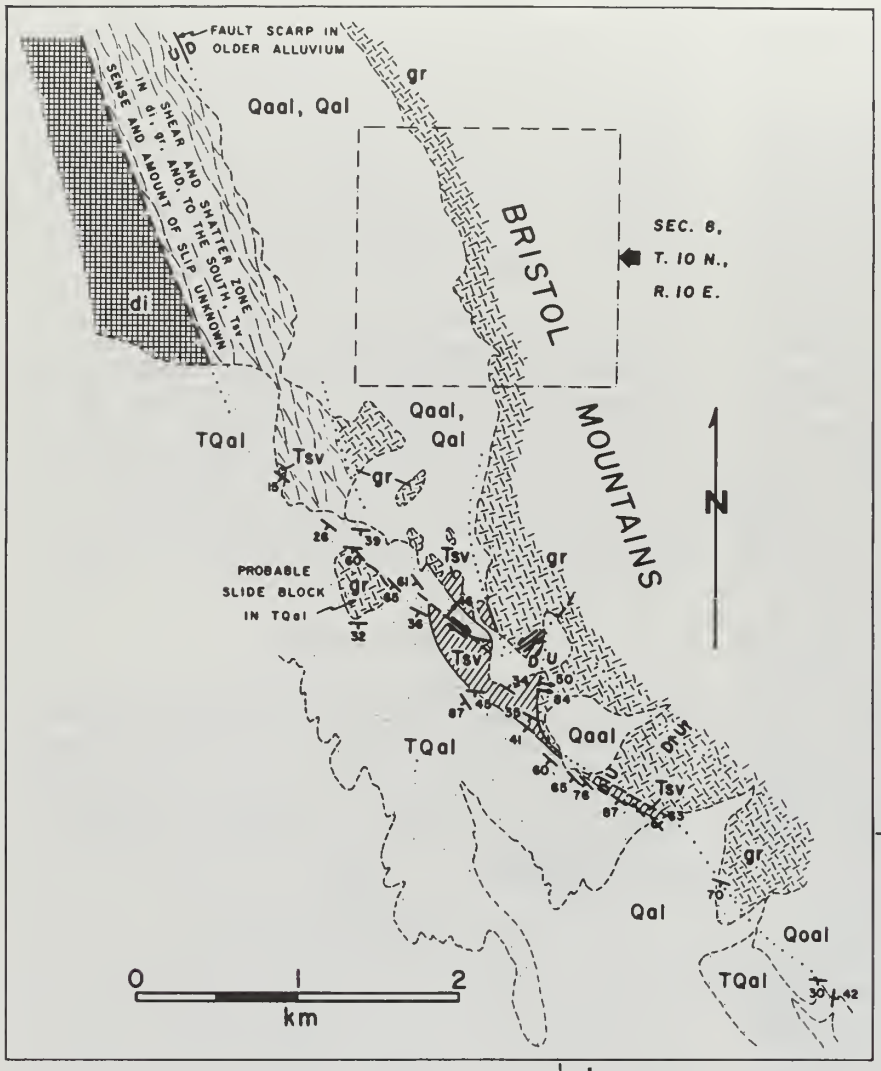

Figure 3. Geologic map of southern Bristal Mountai Broadwell Lake 15' quadrangle. Rock units: di = di rite; gr = undifferentiated granitic racks; Tsv = undi ferentiated Tertiary sedimentary and volcanic roc TQal = late Tertiary ar early Quaternary fanglamera Qaal = Quaternary alder alluvium; Qal = Quaterna alluvium.

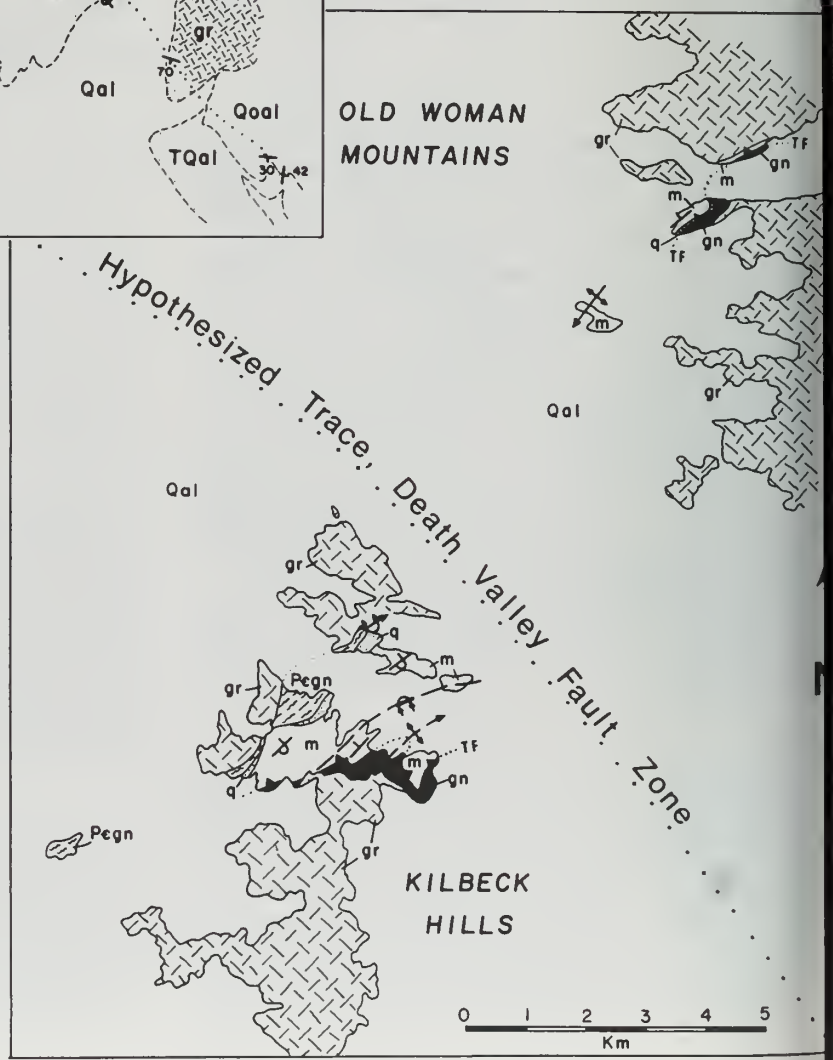

Figure 4. Generalized geologic map af Kilbeck Hills-Old Waman Mauntains area, after Janes (1973) and Evensan (1973). Rock units: Pegn = Precambrian gneiss; q = quartzite; m = marble; gn = gneiss af uncertain age; gr = undifferentiated granitic rocks; Qal = Quaternary alluvium. TF = thrust fault.

oodward-Clyde Consultants who assisted the writer in the ld. Gail Hunt of Southern California Edison's Geotechnical roup encouraged publication of this report. Its content and le have been much improved by the critical reviews of John Crowell, Mason L. Hill, and Bennie W. Troxel.

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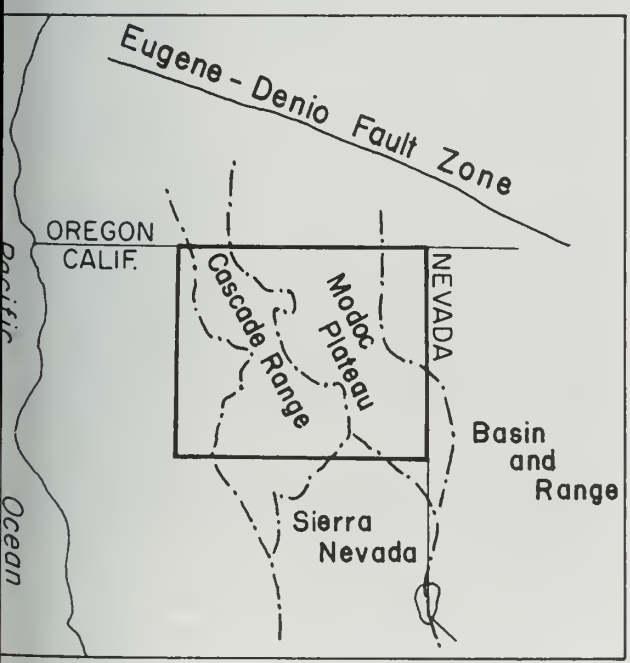
ECTONIC SETTING OF THE MODOC REGION, ORTHEASTERN CALIFORNIA [1]

JUDITH L. HANNAH [2]

The Modoc region of northeastern California (figures 1 and has been largely overlooked in most plate tectonic interpretaons of the western United States. The geology of neighboring rains suggests several possible tectonic models to account for uctures and rock types observed in the Modoc region. Rocks the area comprise interbedded pyroclastic debris, basaltic and desitic flows, mudflows and breccia, and lacustrine and fluvial diments of volcanic origin. The term "plateau," commonly plied to the Modoc terrain, is misleading, as wide-spread block ulting has produced rugged topography similar to that of the sin and Range province, but many of the structural features ve been masked by concurrent volcanism.

The most complete map of the Modoc area is the 1:250,000 ale Geologic Map of California (Alturas and Westwood eets), interpreted from air photographs with field checks (Gay d Aune, 1958; Lydon and others, 1960). The geology of the odoc Lava-Bed quadrangle (scale 1:500,000), including the rthwestern portion of the plateau, was mapped by Powers 932), who included petrographic studies and most of the avail-le chemical analyses of the volcanic rocks of the Modoc Plaau. Two other chemical analyses, not substantially different, e given by Kuno (1965) and Yoder and Tilley (1962). Parts the area, near Alturas and Bieber, have been mapped on a ger scale for hydrological studies by Ford and others (1963). ther work by Anderson (1933 and 1940) and Eichelberger 975) is in the Medicine Lake Highlands area, west of the main ateau.



gure 1. Locotion of the Modoc region and surrounding provinces. Inset ows locotion of map in figure 2.

Reviewed by Robert A. Matthews, E. M. Moores, Thomas E. Gay Jr., and Quintin A. Aune.

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GENERAL GEOLOGY

Exposures of structures within the Modoc Plateau are restricted by the extensive flows of Pleistocene and Holocene basalt, and the lack of major topographic features. Exposed units are predominantly Neogene volcanic rocks and volcanogenic sedimentary rocks. Correlations can be made, however, with units exposed in the Warner Range east of the Modoc region.

Late Oligocene to Miocene rocks of the Cedarville Series include mostly andesitic flows, tuff and mudflow breccia beds, with minor basalt flows and rhyolite tuff. These calc-alkaline rocks were gently folded and faulted near the end of Miocene time (Ford and others, 1963). Within the Warner Range, the type Warner Basalt conformably overlies the Cedarville Series (Russell, 1928), and is therefore dated as late Miocene or younger. The Warner Basalt is moderately coarse grained, is high in alumina with pale yellow olivine phenocrysts, and, in many exposures, exhibits diktytaxitic texture. Since Russell introduced the name, it has been extended to include Pliocene basalt in the northern Sierra Nevada (Durrell, 1959), most of the basalt flows of the Modoc region, and other flows as far north as Bend, Oregon (Powers, 1932). As presently used, "Warner Basalt" is a collective term for petrographically and structurally related rocks without implying contemporaneity.

West of the Warner Range, the Cedarville Series is unconformably overlain by the Plio-Pleistocene Alturas Formation, which is largely made up of lacustrine deposits comprising diatomite, tuffaceous siltstone, and interbedded ignimbrite and olivine basalt flows. Farther south, late Pliocene andesitic volcanic rocks of the Lassen Peak area give way eastward to basalt. Many small basalt shields, each less than 5 miles in diameter, are built up on the plains west of Honey Lake and follow the eastern limit of the Cascade Range northwest to Fall River Mills. Pleistocene and Holocene rocks are predominantly broad flows of high-alumina basalt, including the Burney, Gardens, and Modoc basalts, all correlated with the Warner Basalt. More detailed lithologic descriptions of all rock units are given by Macdonald (1966). The stratigraphic validity of most of these units is highly questionable; more detailed age data and careful mapping are needed to establish formation names and reliable correlations.

The predominant mode of tectonism since Pliocene time has been normal faulting (figure 3). The anomalous Likely fault, reported to have right-lateral, strike-slip displacement (T.E. Gay Jr. and Q. A. Aune, California Division of Mines and Geology, personal communications) trends N35°W for 50 miles across the Modoc Plateau from Devils Garden to the Madeline Plains. Northeast of the Likely fault, most normal faults trend approximately north (e.g., bounding faults of Surprise Valley, Goose Lake, and Tule Lake). To the southwest, most other faults follow the northwesterly trend of the Likely fault but still have predominant components of dip-slip offset. Shawe (1965), however, cited evidence from first motion studies, historic surface faulting, and offset of Cenozoic features, for significant strike-slip components on faults throughout Nevada. This may be true for many of the apparently normal faults of the Modoc Plateau, but determination of total displacement requires more detailed work.

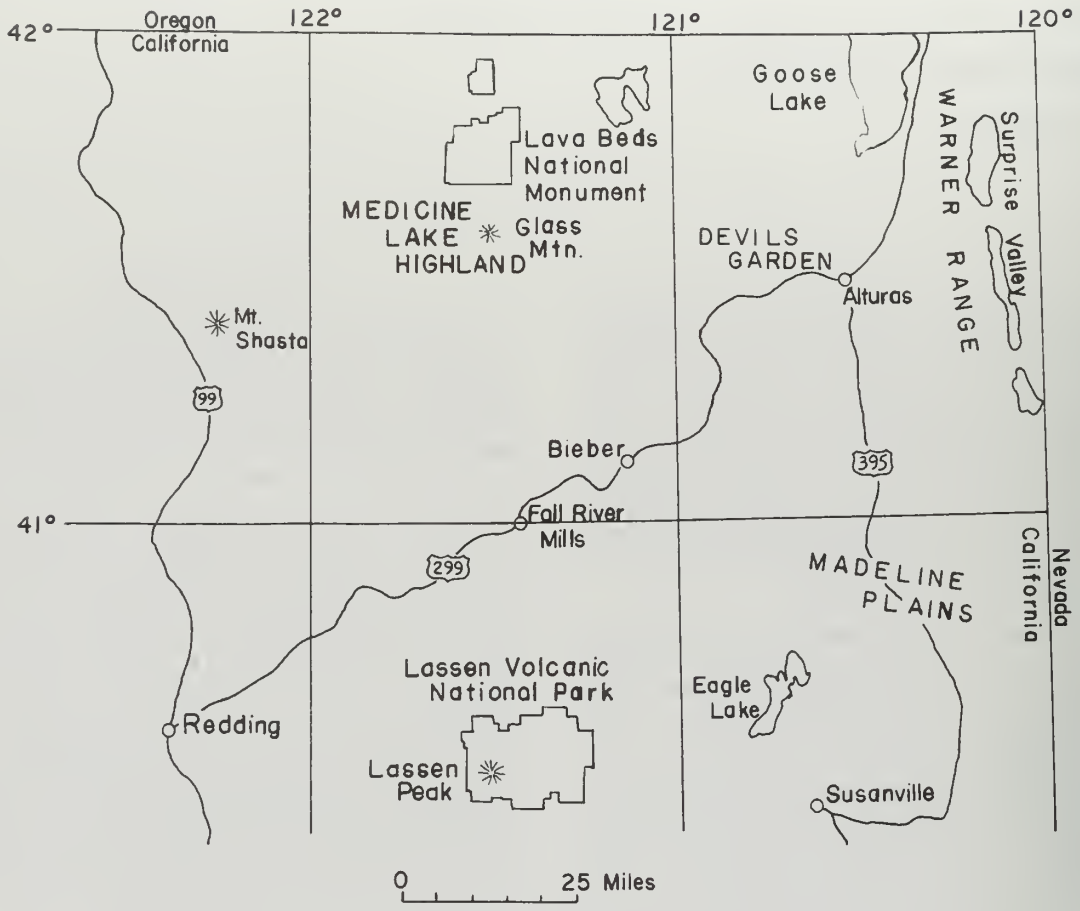

Figure 2. Index map of geographic names used in text.

The onset of block faulting apparently coincides with early extrusions of the basalt (Christiansen and Lipman, 1972; Hamilton, 1975; McKee and Silberman, 1975; Snyder, Dickinson, and Silberman, 1975); flows partially flooded the grabens, and older flows are more commonly faulted. This period also marks the termination of calc-alkaline volcanism (mainly andesitic) and the beginning of bimodal volcanism (predominantly basaltic with minor rhyolitic eruptions). Christiansen and Lipman (1972) placed the transition between tectonic styles in late Miocene time, approximately 8 million years ago.

Despite large offsets on numerous faults, there has been no recorded seismicity in the Modoc region. Although Christiansen and Lipman (1972) stated that areas of basaltic volcanism are seismically active, historic records show few persons felt or recorded earthquakes in the Modoc Plateau. Suppe and others (1975) noted the distinct lack of seismicity in the basaltic plateaus of the northwestern United States, the Snake River Plain, and the volcanic chain traversing New Mexico. Their statements are supported by other surveys (for example, Smith and Sbar, 1974, and Pennington and others, 1974). I have analyzed months of seismic records from the Modoc and Oregon plateaus and the Snake River Plain, which revealed no activity. The failure to record seismicity may result from the lack of permanent seismic stations and inadequate recording intervals on portable units. The apparent aseismicity of the basalt plateaus warrants further study.

TECTONIC AFFINITIES

The Modoc region is bounded on the east and southeast by the Basin and Range Province, on the west by the Cascade Range, and on the north by the structurally and lithologically similar Oregon Plateau. Each boundary is arbitrary, as lava in t Modoc Plateau overlaps the rock units and structures of t surrounding regions. Affinities with each of these areas are co sidered before inferring the Modoc region's present tectonic se ting.

Although the fault block terrain of the Modoc region has be modified by concurrent volcanism, it is related structurally that of the Basin and Range. The Likely and Honey Lake fau lie on strike with the Walker Lane, a zone of right-lateral she extending through western Nevada (figure 3). The Walker La separates the north-trending fault block ranges of central N vada from predominant northwest-trending fault block ranges eastern California. Thus, the Likely and Honey Lake faults a correlative with the Walker Lane. The Walker Lane may ha been an important structure as early as Cretaceous time and still active. Nielsen (1965) described the 1932 Cedar Mounta earthquake, in which right-lateral faulting occurred, and cit numerous examples of offset Tertiary units. Strike-slip faultin related to Basin and Range structures has been described b other authors (Donath, 1962; Wise, 1963; Shawe, 1965; Pea 1969; Hardyman and others, 1975; Lawrence, 1976). Therefor it is reasonable to extend the zone of shearing across northeas ern California.

The Modoc Plateau has lithological and chemical affiniti with the Cascade Range, a volcanic arc complex. The Medici Lake Highland is a calc-alkaline complex forming an eastwar extension of the Cascade Range toward the Modoc Plateau. Th most recent volcanism near the highland resulted in the hig alumina Modoc Basalt of the Modoc Lava Beds National Mon ment, which is chemically similar to the older Warner Basa (table 1). A line of small basaltic shield volcanoes extendin

ble 1. Chemical analyses of basalts from the Modoc Plateau (Powers, 1932). Powers' sample numbers are given.

	Warner Basalt (19123)	*Modoc Basalt (19142)*
.....	47.10	48.98
O_3	18.52	18.92
.....	0.90	1.16
O_3	tr	2.22
.....	7.91	7.12
O	tr	0.09
O	10.89	7.43
.....	11.98	10.04
O	2.33	3.04
.....	tr	0.44
O^+	0.10	0.34
O^-	0.18	0.05
.....	0.09	0.14
al	100.00	99.97

n Honey Lake to Fall River Mills marks the eastern limit of andesitic volcanism of the Cascade Range and the transition predominantly basaltic volcanism. Kuno (1966) described variation in composition of basalt across the Japan arc, from eiite near the trench to high-alumina basalt behind the volic front. This corresponds well to basalt of the Cascade Range of the Modoc and Oregon Plateaus. The volume of basalt, vever, decreases with distance from the Japan trench, in diopposition to the distribution in the northwestern United es. As an alternative model, the extensive plateau basalt may elated to behind-the-arc spreading. The mechanisms operatin behind-arc basins are poorly understood, and the high mina content of the rocks is not readily explained.

The division between the Modoc and Oregon Plateaus is largeased on a political boundary. Donath (1962) described norl faults and possible right-lateral shear in the Oregon Plateau ilar to structures in the Modoc area. The normal faults occur conjugate pairs, however, with trends of N35°W and N20°E ner than north. Powers (1932) correlated basalt of the Ore- Plateau with the Warner Basalt, noting: ". . . as far north Bend, Oregon, the rock is very constant in composition, form occurrence and geological relations." The major change in ology and structure occurs at the boundary with the Cohbia Plateau in northern Oregon, beyond which tholeiitic ba- predominates.

TECTONIC MODELS

A simple tectonic model to explain the wide-spread, highmina basalt flows of the northwestern United States is not mediately apparent. Fault patterns may be explained by eastst extension associated with strike-slip motion between the nerican and Pacific plates. Lithologic variations, however, are so easily understood. Lawrence (1976) did not consider trology in his model. Christiansen and Lipman (1972) simply sented their correlations of age and tectonic style with rock e without interpretations.

Structural patterns may be explained by the relative motion of American and Pacific plates. In late Oligocene time, the ndocino Fracture Zone, the East Pacific Rise, and the Ameri- Trench collided, forming an unstable junction of four plates gure 4). Subsequently, two stable triple junctions formed and grated northwest and southeast to their present positions off int Arena and in the Gulf of California. The present plate undaries involve right-lateral strike-slip motion along the San ndreas fault and possible slow subduction under the Cascade nge. McKenzie and Julian (1971) found that negative travelne residuals for P-wave arrivals at stations in the western nited States are explained by a high-velocity slab dipping 50° to the east under the Cascade Range. The anomalous arrival times are recorded as far south as Mt. Shasta; therefore, subduction is either active or recently terminated just west of the Modoc Plateau.

The total relative motion between the American and Pacific plates cannot be accounted for by measured strain along the San Andreas fault. Atwater and Molnar (1973) deduced from sea floor spreading in the Atlantic, Indian, and South Pacific oceans that the total relative motion between the plates in the last 21 million years has been about 600 km. Estimates of total displacement along the San Andreas fault fall short of this value by almost 300 km. Some of this discrepancy can be accounted for by offset within the Salinian block, but measurements in the western Basin and Range indicate about 100 km of right-lateral displacement (Hamilton and Myers, 1966; Thompson and Burke, 1973). Residual strain may be taken up by a broad zone of right-lateral shear involving the entire Basin and Range. This shear zone requires a strain ellipsoid with maximum principal strain oriented east-west (figure 5); if the maximum principal stress is oriented vertically, the observed pattern of north-trending normal faults results. Wright (1976) addresses the idea of right lateral mega-shear in the Basin and Range, and proposes a new model based on westward displacement of the Sierra Nevada block. He does not suggest a cause, however, for the westward drift.

In south-central Oregon, the dominant structural pattern is conjugate normal faults oriented at about N35°W and N20°E. Assuming the maximum principal stress is oriented north-south, minimum stress east-west, and intermediate stress vertical, the Coulomb-Mohr theory of failure predicts strike-slip faults in the observed pattern. With the northward migration of the Mendocino Triple Junction, east-west extension and vertical stress have become more important. The existing fractures, however, must be at a high angle to those required by a new stress field if a new fracture system is to form. Therefore, recent strain has been taken up by vertical displacement along the old strike-slip fractures.

The structures of the Modoc region are more like those in Nevada than those in Oregon, although minor conjugate fault pairs are present. Lawrence (1976) pointed out that the Eugene-Denio fault zone marks the southern boundary of the rhombic fault pattern described by Donath (1962). To the north, faulting terminates completely, and the Columbia Plateau is deformed by gentle west-trending folds. Thus, the marked east-west extension in the Basin and Range dies out to the north and north-south compression dominates. The change in stress distribution begins just north of a line through the Mendocino Triple Junction and perpendicular to the subduction zone, where the conversion of the American plate boundary from trench to transform has not yet occurred.

To summarize, in central Nevada, residual strain produces right-lateral shear and strong east-west extension, resulting in wide-spread north-trending normal faults. The faulting began when the Pacific and American plates collided about 29 million years ago. In the Modoc region and northern Nevada, the conversion of the plate boundary is more recent (about 8 million years ago) and normal faults are less well developed. North of the Eugene-Denio fault zone, the stress change is very recent, and normal faulting is restricted to movement along the older conjugate shears. Still farther north, there is no east-west extension, and folds, rather than faults, are the dominant structure. Major zones of right-lateral shear may be, in part, tear faults produced by progressive northward decrease in east-west extension.

The extensive basalt flows cannot be explained without further work. Volcanism is a natural consequence of extension and crus-

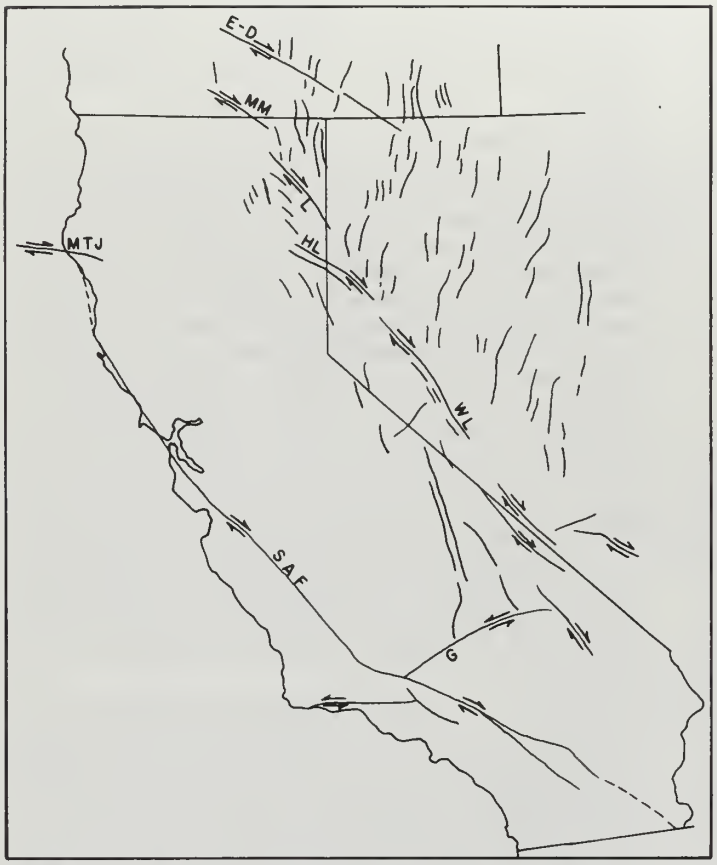

Figure 3. Major Tertiary faults and fault zones in California, Oregon, and Nevada. Lettered features are: (MTJ) Mendocino Triple Junction, (SAF) San Andreas fault, (E-D) Eugene-Denio fault zone, (MM) Mount McLaughlin fault zone, (L) Likely fault, (HL) Honey Lake fault, (WL) Walker Lane, (G) Garlock fault.

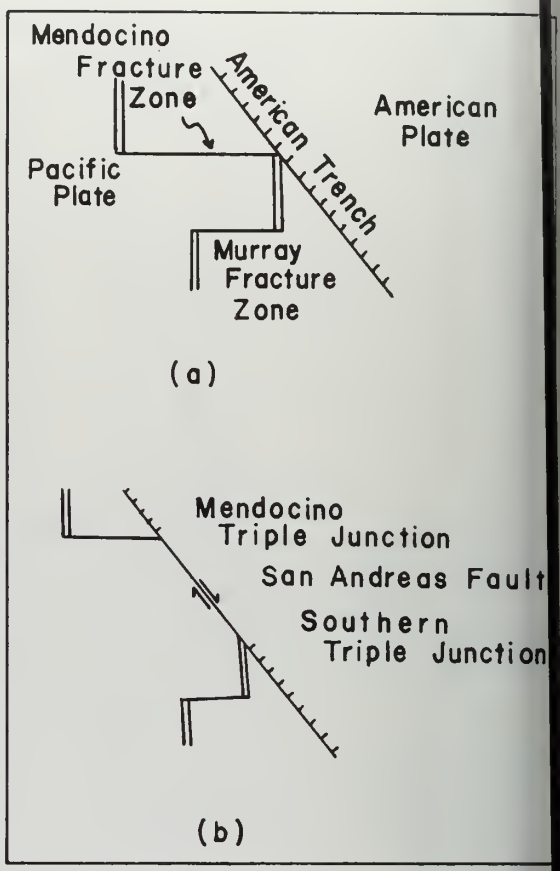

Figure 4. Development of the North American continental mar
(*after Christiansen and Lipman, 1972*). (a) Initial ridge-trench co
sian. (b) Evolution of transform fault and triple junctions.

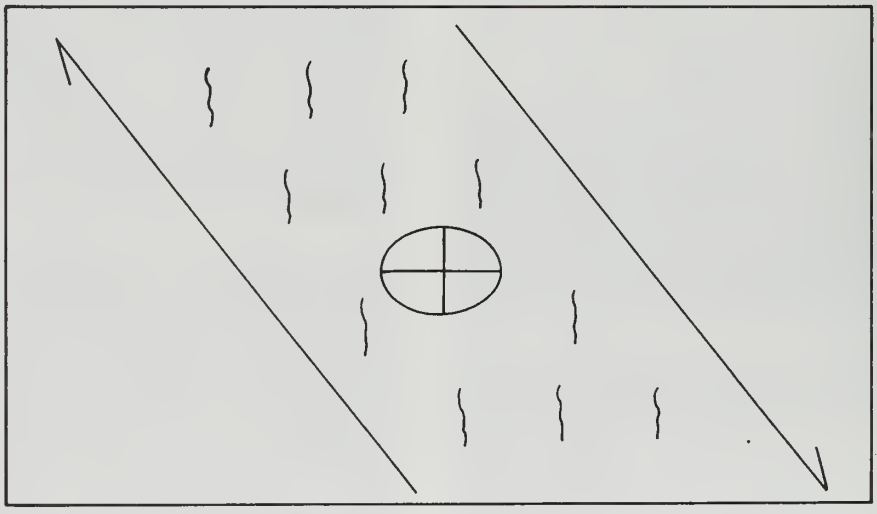

Figure 5. Broad zone of right-lateral shear, required strain ellipsoid, and resultant fracture pattern.

thinning, and is common in the Basin and Range. Suppe and ers (1975) proposed several zones of rifting in areas of recent lcanism where extension is concentrated. The more continu- s flows of the basalt plateaus may be a result of their position hind the Cascade arc. Past migration of the arc westward uld further enhance the high heat flow under the plateaus; idence of westward migration, however, has not been cited. nfortunately, the presence of high-alumina basalt is not under- od. It may be a result of greater depth of melting (Ringwood, 75), but the mechanism is not apparent. Development of a hind arc basin may be important, but its implications are not own.

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AN JUAN-ST. FRANCIS FAULT—HYPOTHESIZED MAJOR MIDDLE TERTIARY GHT-LATERAL FAULT IN CENTRAL AND SOUTHERN CALIFORNIA [1]

DREW P. SMITH [2]

ABSTRACT

The San Juan, Chimeneas, and Morales faults in the sauthern Caast nges, the St. Francis fault in the narthern Transverse Ranges, and the emens Well fault in the Salton Sea regian tagether may have formed one ntinuous northwest-trending fault in central and sauthern California during gocene and early Miocene time. At least 175 km of right-lateral movement curred along this fault, herein called the San Juan-St. Francis fault, before vos obliquely offset by younger cross-cutting faults. A large quartz manzo- e pluton which lies at the northern end of the La Panza Range in the uthern Coast Ranges was displaced fram basement rock of matching litholy in the southeasternmost Little San Bernardino Mountains, north of the ltan Sea. The approximate cumulative 450 km of separation between se twa localities has resulted from incremental offsets approximating: 175 of right-lateral displacement on the ancestral San Juan-St. Francis fault; bsequent left-lateral offset of about 20 km along the San Francisquito ult; followed by 55 km of right-lateral separation along the San Gabriel ult; and most recently by 225 km of right-lateral movement along the San dreas fault. The San Juan-St. Francis fault may have accounted for a bstantial part of the "missing displacement" within the San Andreas fault stem in southern California.

INTRODUCTION

The San Juan-St. Francis fault and the palinspastic maps prented herein are postulated on the basis of four lines of evidence:

(1) The similarity of the quartz monzonite in the northern La Panza Range a body of quartz monzonite in the Thermal Canyon area, southeasternmost ttle San Bernardino Mountains, directly north of the Mecca Hills. It is ggested that both rock bodies were originally part of the same pluton. This rock is informally called the La Panza quartz monzonite in this report.)

(2) The occurrence of a sedimentary megabreccia in the Charlie Canyon rea of northwestern Los Angeles County that appears to have been derived om the La Panza quartz monzonite.

(3) The presence, in the Charlie Canyon area, of a short segment of a reviously unrecognized major east-west trending lateral fault (herein amed the St. Francis fault because of its praximity to the ill-fated St. Francis om).

(4) The dissimilarity of pre-Miocene rocks along opposing sides of the Chimeneas, Morales, St. Francis, and Clemens Well faults, strongly suggesting major lateral displacement along these faults.

The first three lines of evidence are based on field observations by the writer, and the fourth line of evidence is based on analysis of the existing literature.

The 450 km separation of the two occurrences of La Panza quartz monzonite is the net result of displacements along the San Andreas fault and at least three other faults. These displacements, from oldest to youngest, occurred along the San Juan-St. Francis fault, the San Francisquito fault, and the San Gabriel fault. Only the San Andreas fault is now a continuous, through-running fault; all of the older faults have been segmented by movements along younger cross-cutting faults. During the time that movement occurred along the San Francisquito fault, the Big Pine fault may have been the southwestward continuation of San Francisquito fault.

This report is primarily concerned with the presentation of evidence for the existence of the San Juan-St. Francis fault, with the determination of which of the presently-known faults constituted that ancient fault zone, and with the determination of the timing and magnitude of cross-cutting of the San Juan-St. Francis fault by younger faults. An additional outcome of this study is the independent determination of the total lateral displacements along the San Gabriel and San Andreas faults. This is based on the amount of oblique separation of the fault segments that formerly constituted the San Juan-St. Francis fault. The values obtained by this writer are in good accord with those postulated by Crowell (1975b), Ehlig and others (1975), and Bohannon (1975) for the San Gabriel fault, and with those of Crowell (1962) and Ehlig and others (1975) for the San Andreas fault.

Figure 1 is a generalized fault map of the region between the northern La Panza Range and the Salton Sea. The figure shows selected faults and geographic features, and also serves as an index map for figures 2, 3, and 4.

Reviewed by Salem J. Rice, Earl W. Hart, and Perry L. Ehlig.
California Division of Mines and Geology, Los Angeles. This investigation was not funded by CDMG; it was carried out on the author's own time and at his own expense.

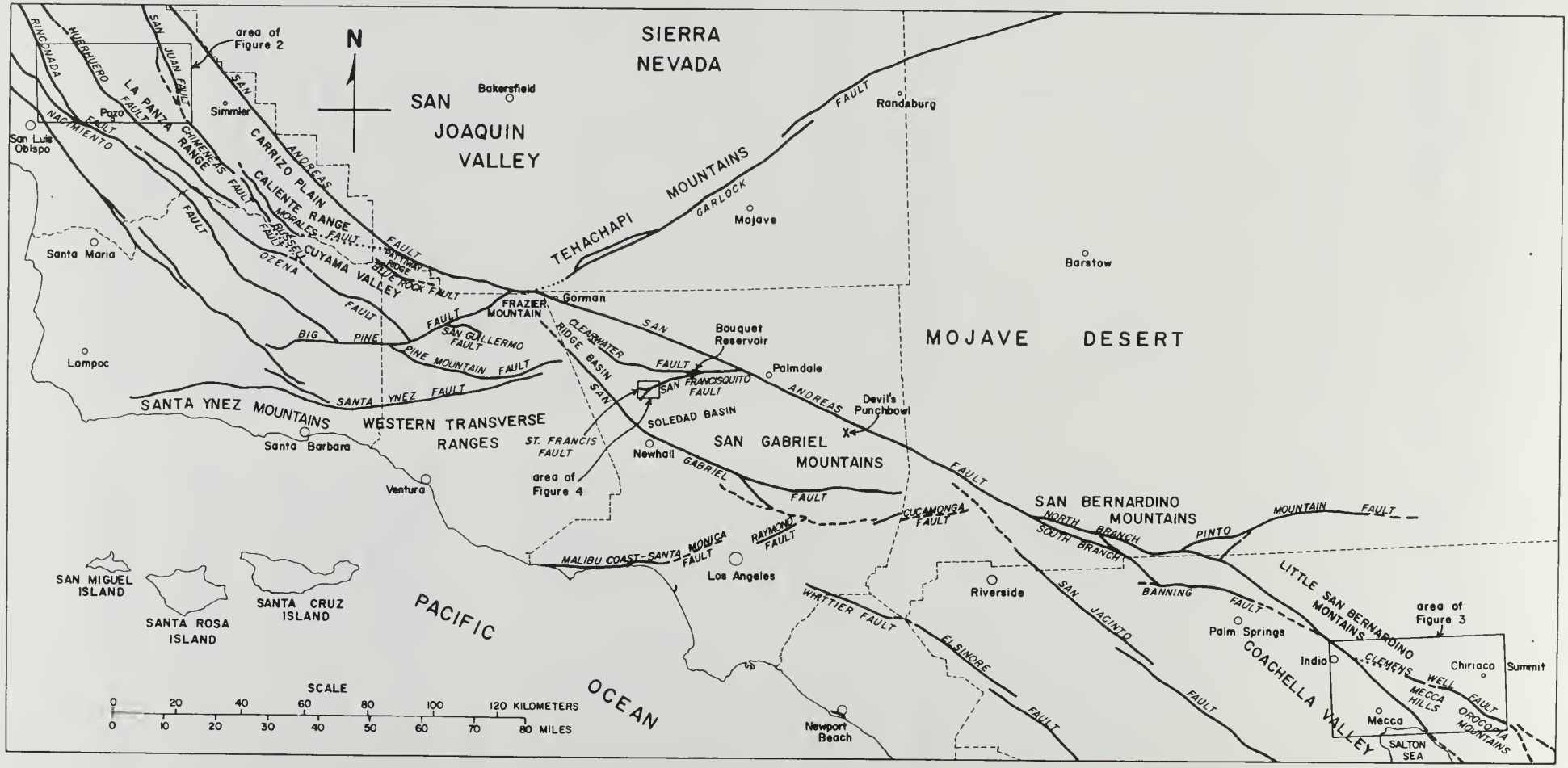
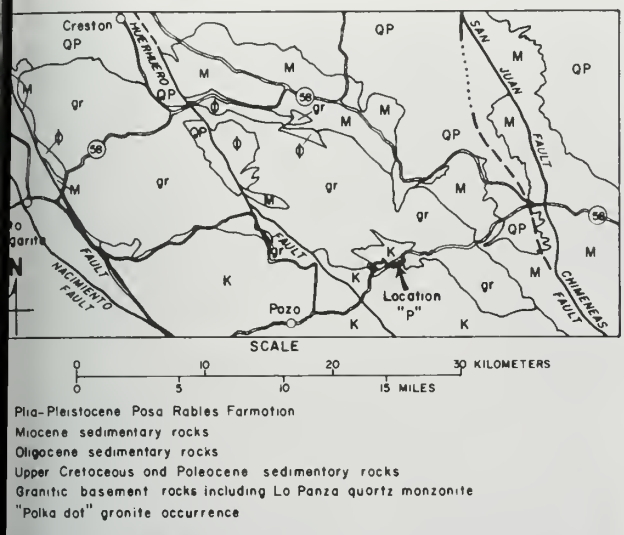


Figure 1. Generalized fault map of southwestern California showing location of the areas shown in figures 2, 3, and 4.

NORTHERN LA PANZA RANGE AREA

Figure 2 shows the geology of the northern La Panza Range a. This map is generalized from Bartow's map (1974).



igure 2. Generalized geologic map of the northern La Panza ange area. Quaternary alluvium omitted; geology not shown outhwest of the Rinconada fault. ***Modified from Bartow (1974).***

The granitic basement exposures in the northern part of the Panza Range occur within a northwest-trending, fault-unded structural block—herein referred to as the La Panza ock. The La Panza block is bounded on the southwest by the nconada fault, and on the northeast by the San Juan-Chime-as fault. Upper Cretaceous and lower Tertiary sedimentary cks overlie the granitic basement rock in the southeastern part the La Panza block, and middle and upper Tertiary sedimen-ry rocks overlie the granitic basement to the northwest. The uerhuero fault (also called the La Panza fault by some work-s) bisects the La Panza block in a northwest-southeast direc-on. It appears, however, that major displacement has not curred along that fault.

More than half of the exposed crystalline basement rock in the rthern La Panza Range is of one lithology—a granitic rock at ranges from quartz monzonite to granodiorite in composi-on. The rock is medium to medium light gray, becoming yel-wish tan where weathered. The texture is distinctly seriate, nging to porphyritic in some facies where phenocrysts of K-ldspar as large as 2 cm occur. Biotite, ranging from tiny flakes masses about 5 mm long, is the principal mafic mineral. This ineral is evenly distributed throughout the rock, and shows a ight alignment, giving some specimens slight foliation. Sphene nd magnitite are the most obvious accessory minerals. The mposition of the rock ranges from 35% to 50% plagioclase, 0% to 30% K-feldspar, 25% to 35% quartz, and 5% to 15% otite. Minor hornblende occurs in some specimens.

Where relatively unweathered outcrops or clasts of this rock ave been exposed to open-air weathering, a distinctive charac-eristic surface appearance has developed. It is the similar dis-nctive appearance of weathered surfaces on rocks in the Charlie anyon area of the northern Transverse Ranges and in the Ther-al Canyon area of the southeasternmost Little San Bernardino lountains that first suggested that some genetic relationship xists among rocks of the three localities.

Ross (1972, p. 24) describes the La Panza Range basement rock, including a description of the microscopic features of the rock and modal analyses plots. The megascopic descriptions he gives for the principal rock type of that area appear to apply equally well to the quartz monzonite of the Thermal Canyon area and the quartz monzonite clasts in the Charlie Canyon megabreccia.

Although the La Panza quartz monzonite is the dominant rock type in the northern La Panza Range area, there are several other smaller intrusives and dikes of different lithologies. No attempt has been made to correlate all of these with rocks in the Charlie Canyon and Thermal Canyon areas, but at least one very distinctive dike rock may be common to both the northern La Panza Range area and the Thermal Canyon area. This rock, which occurs at location "P" shown on figure 2, contains very conspicuous spherical alterations, from 2 cm to 4 cm in diameter, that were apparently caused by reaction of the magma with small inclusions. Rocks of very similar appearance, called "polka dot" granite (Pelka, 1971), occur as cobbles and boulders in the Punchbowl Formation at Devil's Punchbowl, and in the lower part of the Ridge Basin Formation (Ehlig, 1976) about 17 km northwest of the Charlie Canyon area. Ehlig (1976) indicates that the "polka dot" granite also occurs in the granitic terrane north of the Mecca Hills.

The "polka dot" granite locality in the La Panza area is very near the contact with Upper Cretaceous and lower Tertiary sedimentary rock. Thus the "polka dot" granite probably has been exhumed only within the last few million years. This locality, therefore, was not likely to have been the source for the "polka dot" granite clasts that occur in Tertiary sedimentary rocks within the Transverse Ranges.

THERMAL CANYON AREA

Figure 3 shows the generalized geology of the region north of the Salton Sea. Crowell (1975a) gives a fairly detailed description of the geology of the central and southeastern parts of this area. The most salient geologic feature is the San Andreas fault zone which separates the Salton trough on the southwest from Tertiary sedimentary rocks and pre-Tertiary crystalline basement rocks on the northeast. The west-northwest trending Clemens Well fault passes along the northeast side of the Orocopia Mountains and is buried beneath Quaternary sediments along the northeast side of the Mecca Hills. The fault presumably converges with the San Andreas fault zone, but the true point of convergence is uncertain. Crowell (1975a, p. 107) indicates that the San Andreas fault zone is quite wide in this area, with abandoned splays of the fault occurring 4 to 8 km northeast of the present active trace. The Hidden Springs fault would appear to be the northeasternmost splay of the San Andreas fault system, but the presence of the same rock type (Orocopia Schist) on both sides of the fault indicates there has been no significant displacement along that fault. The true zone of major displacement probably does not extend more than 5 km to the northeast of the present active San Andreas fault trace. Therefore, the writer tentatively assumes that the Clemens Well fault converges with the San Andreas fault zone somewhere between Thermal Canyon and Fargo Canyon.

The Clemens Well fault generally forms the boundary between two distinctly different suites of basement rock. To the southwest, in the Orocopia Mountains (Crowell, 1975) and the Mecca Hills (Sylvester and Smith, 1975), the basement consists of the

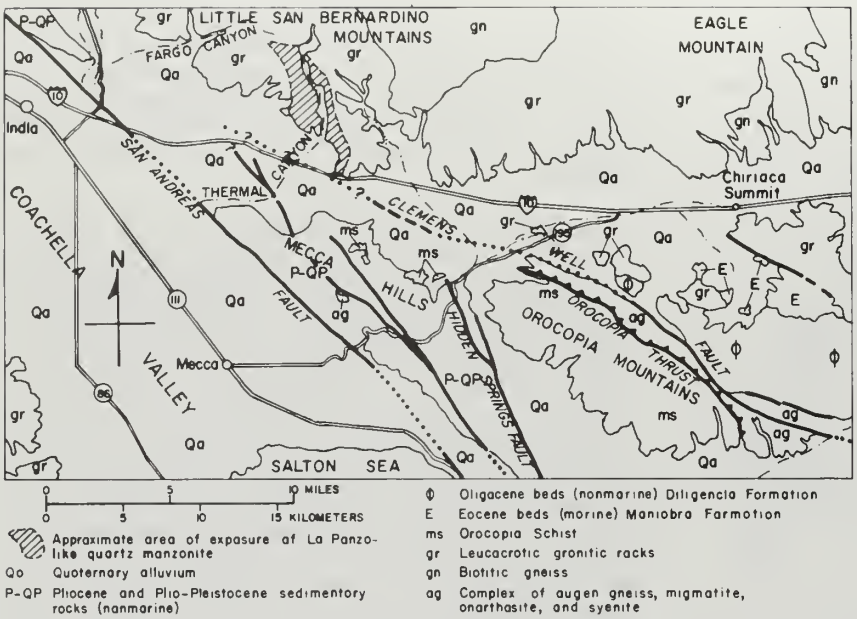


Figure 3. Generalized gealagic map af the reg narth af the Saltan Sea. *Modified from Ro (1965) and Jennings (1967).*

Orocopia Schist, augen gneiss, migmatite, blue-quartz gneiss, anorthosite and syenite. To the northeast the basement consists primarily of leucocratic granitic rocks, with biotite-rich gneisses becoming common farther to the north and northwest.

The cross-hatched area in figure 3 along the upper part of Thermal Canyon indicates the approximate area of the observed occurrence of a type of quartz monzonite that appears to be identical to the La Panza quartz monzonite. This occurrence lies to the northeast of the northwest projection of the Clemens Well fault.

Ehlig (1976) states that the "polka dot" granite occurs in a pluton along the northern margin of the Orocopia Mountains and the Mecca Hills. At this time, the writer has ascertained only the existence of, and delineated the approximate boundaries of the La Panza-type quartz monzonite; the writer did not look for or observe occurrences of the "polka dot" granite. If the "polka dot" granite that Ehlig (1976) refers to is the same as that which occurs in the northern La Panza area, and it occurs in a similar manner—as small intrusive bodies and dikes within a larger pluton, then an even stronger case can be made for the rock in the Thermal Canyon area to have once been contiguous with the basement rock of the northern La Panza Range.

CHARLIE CANYON AREA

Figure 4 shows the general distribution of faults and rock types in the Charlie Canyon area. Relatively detailed mapping of the area has been done by Szatai (1961), Sams (1964), and Konigsberg (1967), all part of unpublished theses. Many of the findings and interpretations of the writer in some parts of the area differ significantly from those of the previous workers.

Briefly, the area contains a 2,200 m thick sequence of coarse south-dipping Oligocene(?) fluvial beds that is bounded on the south by the St. Francis fault zone. Both the St. Francis fault zone and the Oligocene strata strike generally east-west, and are truncated on the east by the San Francisquito fault and on the west by the Camp fault. The area southeast of the San Francisquito fault is underlain mainly by Pelona Schist, with Miocene sediments, including the nonmarine Mint Canyon Formation lapping onto the schist from the south. Upper Miocene strata including the Mint Canyon Formation and the marine Castaic Formation, are downfaulted against the Oligocene strata and pre-Tertiary crystalline rocks along the Camp fault. To the north, marine strata of Paleocene or Eocene age are thrust over the Oligocene strata along the Bee Canyon fault.

Two geologic features of this area are of major importance to this discussion: (1) a coarse, angular debris-flow type of sedi

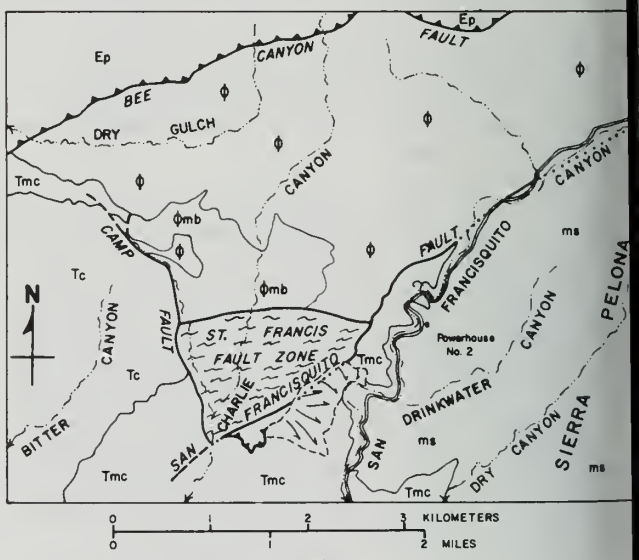


Figure 4. Generalized gealagic map of the Charlie Canyon Area. Quaternary alluvium amitted and lacal expasures of basement rock beneath aligacene beds and Charlie Canyan megabreccia nat shawn. *Modified from Sams (1964).*

ntary breccia, herein referred to as the Charlie Canyon mega-ccia, which occurs at the top of the sequence of Oligocene ata and (2) the wide east-west trending fault zone herein lled the St. Francis fault.

The megabreccia is the unit mapped as "quartz diorite mega-ccia" of the Vasquez Formation by Sams (1964). It occurs massive beds, many at least 10 m thick, containing completely sorted angular granitic debris. The clasts range to as much as m in size, with the 10cm to 50cm size range being dominant. is material is usually enclosed in a dark gray muddy matrix. me parts of the megabreccia are very well cemented and crop t boldly. The megabreccia obtains a maximum thickness of out 500 m (Sams, 1964) and is exposed over an area of about square kilometers. It occurs immediately adjacent to the rthern margin of the St. Francis fault zone, but not farther uth, within the zone.

The age of the megabreccia is not well known. All workers in is area have considered this sequence of fluvial beds in Charlie anyon to be of Oligocene age—primarily on the basis of their nilarity to other Oligocene nonmarine strata in this region, ch as occur in the Sespe, Vasquez, and Simmler Formations. uehlberger (1958), Sams (1964), and Konigsberg (1967) ave, in fact, referred these beds to the Vasquez Formation. The harlie Canyon beds, however, have not been dated by paleonto-gical or radiometric methods. The contact relationships in this ea do not add much information. Scattered exposures show the harlie Canyon beds lying unconformably on granitic and gneis-c basement rock. Upper Miocene strata unconformably overlie e Charlie Canyon beds on the northwest. The Charlie Canyon eds are in fault contact with all other rock units in the area. It ems probable, however, that the megabreccia, which is at the p of the Charlie Canyon sequence, was deposited some time etween middle Oligocene and early Miocene time.

A very important characteristic of the megabreccia is that out half of the exposures, including all exposures within 500 eters of the St. Francis fault zone, are composed almost exclu-vely of debris that appears to be identical to the La Panza uartz monzonite. The more northerly exposures of the mega-reccia, including the very prominent inselberg on the west side f Charlie Canyon, are composed mainly of quartz diorite debris. ock of similar composition and appearance is exposed in much f the southeastern part of the northern La Panza Range crystal-ne basement area. The quartz diorite there, however, is nearly verywhere severely weathered, so the true extent of its similarity o the quartz diorite clasts in the Charlie Canyon megabreccia as not been determined. No "polka dot" granite clasts have een found within the Charlie Canyon megabreccia.

The St. Francis fault zone is nearly 800 m wide and contains lternating slices of severely crushed granitic, gneissic, and schis-se rock, with some lenses of coarse sedimentary debris occur-ing between the slices. The gross fabric of the fault zone is either ertical or steeply south dipping. A number of dike-like zones of emented tectonic breccia occur along the northern margin of he fault zone. These tectonic breccias are of three separate types: 1) chert and jasper fragments cemented by silica or limonite, 2) serpentinite fragments cemented by what appears to be mag-nesite, and (3) limestone fragments cemented by calcite. All of hese breccias show evidence of repeated episodes of shattering nd recementing. The breccia zones are steeply-dipping, irregu-ar to lenticular in plan, roughly oriented east-west in their long-st direction, and are as much as 25 m thick and 150 m in length.

The fault zone is partly covered, especially on the west side of Charlie Canyon, by coarse, angular gneiss and schist breccia. Charlie Canyon, which trends north-south through the east-west trending fault zone, exposes the anatomy of this fault zone in an excellent manner.

The relationship between the San Francisquito and St. Francis faults, as shown in figure 4, differs significantly from the mapping of Sams (1964) and Konigsberg (1967). Sams and Konigsberg failed to recognize the St. Francis fault zone for what it is. They do show an east-west trending fault along the northern margin of the zone, which they consider to be a westward-curving branch of the southwest-trending San Francisquito fault. The St. Francis fault zone is, however, truncated on the east by the San Francisquito fault as shown. The San Francisquito fault continues to the southwest, as shown, but is partially obscurred by massive landsliding on the west side of San Francisquito Canyon. The San Francisquito fault truncates the Mint Canyon Formation on the northwest, and, on that basis, the fault can be located and followed to the southwest of the landslide area. The Camp fault terminates against the San Francisquito fault, and the San Francisquito fault continues on a fairly straight course to the southwest. It can be traced for nearly 1 km within the Mint Canyon Formation beds as a zone of disrupted bedding and small faults.

Sams (1964) shows the megabreccia unit as also occurring within the area of the St. Francis fault zone. These occurrences, however, are not part of the Charlie Canyon megabreccia which, as herein defined, occurs only to the north of the fault zone. Although they are similar in gross appearance, the sedimentary breccia overlying the fault zone is composed either entirely of gneissic debris or entirely of Pelona Schist debris. The megabreccia to the north of the fault zone contains only rare gneissic clasts and no Pelona Schist clasts.

In summary, the exposures in Charlie Canyon show a short segment of a major fault zone, the eastern and western continuations of which have now been displaced by younger cross-cutting faults. The thick sequence of highly sheared and crushed slices of various rock types, including types exotic to this area such as serpentinite, indicates a long history of lateral displacement along this fault zone. It is very unlikely that this fault zone is merely part of the San Francisquito fault. The San Francisquito fault is commonly well defined, especially where it is exposed to the northeast of the St. Francis fault. The total apparent zone of shearing and deformation along that part of the fault is, in places, less than 10 m thick.

There is no apparent nearby source for the quartz monzonite that constitutes the Charlie Canyon megabreccia. The granitic and gneissic basement that underlies the Oligocene strata to the north of the St. Francis fault zone is of different lithology than the megabreccia clasts. The distinctive Pelona Schist underlies the area to the southeast of the San Francisquito fault. The slices of rock within the St. Francis fault zone include Pelona Schist, gneisses, and some granitic rock, but not the types that constitute the megabreccia. Sams (1964) and Konigsberg (1967) suggest that the megabreccia was derived from a granitic body to the southeast that formerly overlay the Pelona Schist, and which has subsequently been completely removed by erosion. The granitic rocks that now overlie the Pelona Schist along the south side of Sierra Pelona are of a different type. Farther south, within the Oligocene, Miocene, and Pliocene sedimentary rocks in the Soledad basin, the writer could find no granitic clasts of the type that

constitute the Charlie Canyon megabreccia, indicating that no such rock was eroded from Sierra Pelona in middle to late Tertiary time.

It is certain, however, that the source rock for the megabreccia was nearby. The angular, debris-flow character of the megabreccia indicates that the transport distance was no more than a few kilometers, and the almost monolithologic composition of parts of the megabreccia indicates that the debris was derived directly from a bedrock source, not from the reworking of older deposits. Searching farther afield in the San Gabriel Mountains and the northern and western Transverse Ranges has yielded no rock of even doubtful similarity to the quartz monzonite of the Charlie Canyon megabreccia. To the knowledge of the writer, the La Panza quartz monzonite is the nearest exposure of rock of similar lithology.

OLIGOCENE AND MIOCENE FAULTING

The determination of the sequence of fault movements that brought about the 450 km of separation between the La Panza and Thermal Canyon localities is a complex problem. The sequence of fault movements proposed by this writer is based on consideration of geometric constraints, and on an analysis of the literature regarding the nature of the many faults in the region between the northern La Panza Range and the Thermal Canyon areas—especially along a number of faults for which lateral displacement has been suggested but not specified.

Some of the most important considerations are:

(1) The separation between the northern La Panza Range and Thermal Canyon areas is in the northwest-southeast direction. The writer presumes that most of this separation was accomplished by movements along northwest-trending right-lateral faults.

(2) Therefore, geometry indicates that the original fault that separated the two localities had to be along the northeast side of the La Panza area and along the southwest side of the Thermal Canyon locality.

(3) The basement rock beneath the Charlie Canyon megabreccia is of different lithology than the megabreccia clasts. Therefore the quartz monzonite body that lay across the St. Francis fault to the south had to already have been displaced some distance along the fault to that position. Thus, the age of the megabreccia indicates a portion of the time during which the St. Francis fault was active. Considering geometric constraints again, the Thermal Canyon locality could not have occupied the position south of the St. Francis fault; the La Panza area could have.

(4) The La Panza area is 200 km from the Charlie Canyon area. The San Andreas fault cannot have accounted for any of this separation since both localities occur well to the southwest of the San Andreas fault. Therefore, most of this separation must have occurred along northwest-trending faults that lie to the southwest of the San Andreas fault.

There are many northwest-trending faults to the southwest of the San Andreas fault in the region between the Charlie Canyon area and the northern La Panza Range. These are shown in Figure 1 and include the San Gabriel, Ozena, Russell, Blue Rock, Morales, Chimeneas, and San Juan faults. In studying the literature regarding these faults, an important consideration is that displacement may have occurred during Oligocene time, and much of it almost certainly occurred before late Miocene time. Thus, evidence for displacement of lower Miocene, Oligocene, or older rocks is important; a lack of evidence for significant displacement of upper Miocene or younger rocks does not eliminate a fault from consideration.

Because of space limitations, a discussion of the geologic character of each of the above-listed faults is not possible; only the significant findings will be discussed. The references used by the writer in the study of each fault are:

San Gabriel: Crowell (1975b); Paschall and Off (1961); Ehlig (1973)
Ozena: Exum (1957); Schwade and others (1958); Payner (1960); Vedder (1968); Bartaw (1974).
Russell: Schwade and others (1958); Bartaw (1974).
Blue Rock: Schwade and others (1958); Frakes (1959).
Morales: Schwade and others (1958); Vedder and Repenning (1965); Bartaw (1974).
Chimeneas: Ross (1972); Bartaw (1974).

All of the evidence suggests that the Chimeneas and Morales faults formed the original westward continuation of the St. Francis fault. The Chimeneas fault bounds the La Panza block on the northeast and separates the granitic basement of the La Panza block from gneissic basement to the northeast (Ross, 1972, p. 28; Bartow, 1974, p. 138). The granitic terrane in the northern La Panza Range almost certainly had to have been carried northwestward along that fault.

To the southeast there is the question of whether the major through-going fault swings eastward along the Morales fault, or continues southeastward along the Russell and Ozena faults (as suggested by Schwade and others, 1958, p. 85; Bartow, 1974, p. 138). These three faults, and in fact all of the above listed faults, show significant contrasts in their pre-Miocene lithology along opposite sides of each fault. In every case, the evidence allows that major lateral displacement of the pre-Miocene rocks could have occurred along each fault. The pre-Miocene lithologic differences along each fault could also have been caused by some sequence of vertical movements. The Morales fault, however, has pre-upper Miocene lithologic differences along the fault that are difficult to explain by vertical movements alone. The Paleocene Pattiway Formation (Hill and others, 1958, p. 2977; Vedder and Repenning, 1965) is of marine origin in occurrences to the north of the Morales fault, and appears to be of nonmarine origin in occurrences along Pattiway Ridge to the south of the fault. These exposures, on opposite sides of the fault, are separated by only several kilometers and attain thicknesses of at least 2,000 m on the north and 600 m on the south. The overlying Oligocene Simmler Formation (Hill and others, 1958, p. 2981) shows even greater differences across the Morales fault in the same area. Nine hundred meters of pebble to boulder conglomerate and conglomeratic sandstone overlie the Pattiway Formation to the south of the fault. The conglomerate contains mainly biotite gneiss and quartz monzonite clasts. A similar thickness of Simmler Formation strata overlies the Pattiway Formation to the north of the fault. These beds, however, are almost entirely sandstone and siltstone, with only a 10 m thick basal conglomerate. Hill and others (1958, p. 2983) state that the basal conglomerate consists of rounded cobbles of granitic, porphyritic, and quartzitic rocks. To the writer, it seems very unlikely that the northern and southern occurrences of this unit could have been originally deposited as one continuous body with such a strong change in lithology occurring over such a short distance.

Schwade and others (1958) studied the subsurface geology along the central part of the Morales fault. They make the following statement (p. 89):

"The Morales fault abruptly separates obviously different pre-Miocene rocks and it is considered likely that a steep lateral fault lies concealed beneath Caliente anticline. This concealed probable lateral fault zone may continue southeastward as the unnamed fault in outcrop which separates the Eocene and Oligocene? sediments from Cretaceous and granitic basement at the eastern end of Cuyama Valley."

The "unnamed fault" at the eastern end of Cuyama Valley is obably the "Blue Rock fault" of Frakes (1959, p. 60). The orales fault, as it is now delineated, is a north-dipping Quaterry thrust fault forming the southern boundary of the Caliente ange. This writer believes that the thrust fault steepens at pth, passing into the older lateral fault, and that fault contins east-southeastward to the San Andreas fault where it is uncated.

The above findings do not eliminate the possibility of the ussell-Ozena fault having been the southeastward continuation the Chimeneas fault. The Russell-Ozena fault, however, prents several major difficulties, especially when attempting to nnect it to the St. Francis fault. The Ozena fault terminates at e Big Pine fault. Hill and Dibblee (1953, p. 452) and Poyner 960, p. 84) suggest that the San Guillermo fault was the ntinuation of the Ozena fault, and that they were subsequently parated by 13 km of left-lateral displacement along the Big ine fault. The San Guillermo fault swings toward the east, but as not been mapped farther east through a region of complex ology south of Frazier Mountain. If this fault is projected ast-southeastward, it would appear to be the possible westward ntinuation of the St. Francis fault. This is very improbable ecause (1) it would mean that there has been no lateral dislacement along the San Gabriel fault since before upper Mioene time and (2) it in no way brings about a better distribution f the basement rocks of the region when a palinspastic map is nade.

The Morales fault, on the other hand, not only has stronger vidence for major lateral displacement, but also swings eastward at a geographic position that makes it very likely to have een the westward continuation of the St. Francis fault. This hoice requires that about 55 km of right-lateral offset has subsequently occurred along the San Gabriel fault. This value is based n the writer's interpretation that the San Gabriel fault coninued on a straight course to the northwest and was later trunated by the San Andreas fault—as opposed to the view that the San Gabriel fault was the earlier southeasterly continuation of an already-existing San Andreas fault. A value of about 80 km would be required if the latter case is true.

The original continuation of the St. Francis fault to the east r east-southeast must also be determined. The structural relationships in the Charlie Canyon area indicate that the St. Francis fault was truncated on the east by lateral movement along the San Francisquito fault. Because the San Francisquito fault has a northeasterly trend, the writer presumes that the movement was in the left-lateral sense. This presumption, however, is apparently supported by provenance studies of the Mint Canyon Formation within San Francisquito Canyon. Clastic material in parts of the Mint Canyon Formation that lie immediately southeast of the San Francisquito fault was derived from sources to the northwest of the fault. These sources have subsequently moved at least a few kilometers left-laterally to the southwest. The geologic relationships along the San Francisquito fault do not indicate the total magnitude of displacement, but it must have been at least 14 km—the distance from the Charlie Canyon area to Bouquet Reservoir—because no faults diverge eastward or southeastward from the San Francisquito fault along that interval. At Bouquet Reservoir the fault history is complicated by the convergence of the right-lateral Clearwater fault with the San Francisquito fault.

The St. Francis fault, however, must have continued to the east or east-southeast from some point on the southeast side of the San Francisquito fault, regardless of what the actual magnitude of displacement was along the San Francisquito fault. This continuation undoubtedly has been obliquely offset to the southeast by the San Andreas fault. Furthermore, the continuation of the St. Francis fault, to the east of the San Andreas fault, must lie to the southwest of the quartz monzonite in the Thermal Canyon area. The Clemens Well fault is the most likely possibility. The necessary contrast in pre-upper Miocene rocks occurs along opposing sides of this fault. Crowell (1975, p. 108) states that the Clemens Well fault is a major strike-slip fault. He also states that it is unlikely that the slip is greater than several tens of kilometers, because clastic debris in the Diligencia Formation (Oligocene or early Miocene age) on the northeast side of the fault is apparently derived from the basement terrane that now lies across the fault to the southwest. This situation, however, places no constraints on major movement having occurred along the Clemens Well fault shortly before the deposition of the Diligencia strata.

PALINSPASTIC MAPS

The palinspastic maps in figure 5 show the proposed sequence of faulting that brought about approximately 450 km of separation between the quartz monzonite bodies in the northern La Panza Range and the Thermal Canyon area. Figure 5A shows the San Juan-St. Francis fault at the time of initiation of major right-lateral displacement. This presumably occurred sometime during the Oligocene Epoch. In figure 5B, about 15 km of offset had occurred, and deposition of the Charlie Canyon megabreccia was taking place. In figure 5C, nearly 175 km of offset had occurred along the San Juan-St. Francis fault. This movement was probably completed by middle Miocene time, but possibly by as early as late Oligocene time. Subsequently the San Juan-St. Francis fault was obliquely offset by between 14 and 24 km of left-lateral movement along the San Francisquito fault, as shown in figure 5D. This movement occurred prior to or during late Miocene time. In this figure the Big Pine fault is shown as being the original southwestward continuation of the San Francisquito fault, although the writer has no direct evidence to support this. Figure 5E shows another new fault, the San Gabriel fault, obliquely offsetting the older faults by about 55 km of right-lateral movement. Geologic relationships along this fault (Crowell, 1975b) indicate that most of the displacement occurred during Pliocene time. Figure 5F shows the fault distribution as it is at the present. The San Gabriel and all older faults have been obliquely offset by about 225 km of right-lateral displacement along the San Andreas fault.

No attempt has been made, in these palinspastic maps, to straighten out the curves in the older faults. They were probably much straighter at the time they were active, but have subsequently undergone deformation. The fact that displacement occurs along the present San Andreas fault, with its bends, indicates that such deformation is a real and ongoing process.

The distribution of Pelona Schist along the San Juan-St. Francis fault is also shown in figure 5—to aid in the discussion that follows.

DISCUSSION

The hypothesized San Juan-St. Francis fault is just that—a hypothesis. The scheme presented here is not without uncertainties and inconsistencies. Of all the evidence presented, the greatest uncertainty lies in the La Panza quartz monzonite

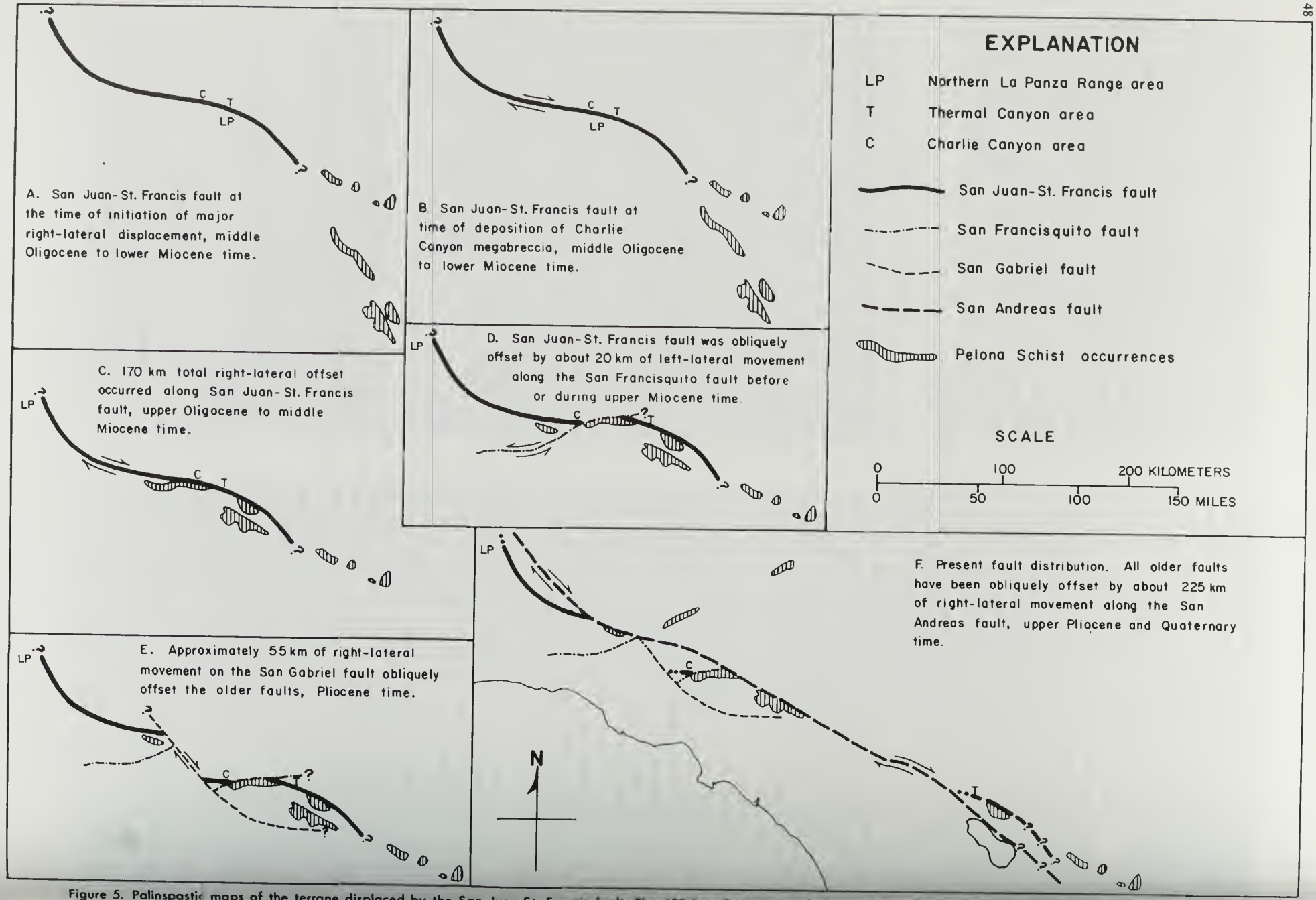

A. San Juan-St. Francis fault at the time of initiation of major right-lateral displacement, middle Oligocene to lower Miocene time.
B. San Juan-St. Francis fault at time of deposition of Charlie Canyon megabreccia, middle Oligocene to lower Miocene time.
C. 170 km total right-lateral offset occurred along San Juan-St. Francis fault, upper Oligocene to middle Miocene time.
D. San Juan-St. Francis fault was obliquely offset by about 20 km of left-lateral movement along the San Francisquito fault before or during upper Miocene time.
E. Approximately 55 km of right-lateral movement on the San Gabriel fault obliquely offset the older faults, Pliocene time.
F. Present fault distribution. All older faults have been obliquely offset by about 225 km of right-lateral movement along the San Andreas fault, upper Pliocene and Quaternary time.
LP
T
C
EXPLANATION
LP Northern La Panza Range area
T Thermal Canyon area
C Charlie Canyon area
San Juan-St. Francis fault
San Francisquito fault
San Gabriel fault
San Andreas fault
Pelona Schist occurrences
SCALE
0 100 200 KILOMETERS
0 50 100 150 MILES
N

Figure 5. Palinspastic maps of the terrane displaced by the San Juan-St. Francis fault. The 450-km of separation

relation. The rocks at the different localities may not have ginally been part of the same body, regardless of how similar y are.

The Blue Rock fault is another problem in this proposal that s not been explained. At least 1,600 m of Paleocene, Eocene, d Oligocene strata occur to the northeast of the fault. South the fault Miocene strata of the Miocene Caliente Formation t directly on granitic basement rock. This relationship may ly be the result of some sequence of vertical movements along fault. It is hazardous, however, to disregard northwest-trend- faults in this region by employing such a convenient assess- nt.

Another problem not considered in this study is the question the northwest continuation of the San Juan-St. Francis fault. he fault appears to continue on a north-northwesterly course the Red Hills fault, but then is buried beneath Plio-Pleistocene posits of the Paso Robles Formation. If projected on the same urse, it would converge with the San Andreas fault in the bsurface. If the terrane that presently occurs to the northeast the San Andreas fault is considered to be "in place," then the rth-northwest continuation of the Red Hills fault would have nverged with that terrane 10 to 20 km southeast of Tejon Pass ring Oligocene and Miocene time. Possibly, then, the San ndreas fault to the northwest of Tejon Pass was the northwest ntinuation of the San Juan-St. Francis fault. If this is the case, en the San Juan-St. Francis fault was truly the ancestral San ndreas fault in southern California.

Pelona Schist occurrences are outlined on figure 5 to show at the inclusion of the San Juan-St. Francis fault into palin- astic maps of central and southern California also brings most f the scattered schist occurrences back into one area in south- asternmost California. It is postulated that the San Juan-St. rancis fault passed through and offset this schist body the same s it did the smaller La Panza quartz monzonite body farther to he north. Pelona-type schist presently occurs in southeastern alifornia in the southeastern Chocolate Mountains and in the astern Picacho area (Haxel and Dillon, 1973). The schist oc- urrences outlined in figure 5 represent the present exposures of he schist and do not imply that it was necessarily exposed at the urface in the past.

One drawback to the credibility of the above line of evidence s that no account is offered for two other widely separated ccurrences of Pelona-type schist: the Randsburg Schist in the orthern Mojave Desert, and the slice of schist along the south- vestern part of the Garlock fault.

Two other unexplained problems are: (1) Where is the ortheastward continuation of the San Francisquito fault and (2) where is the northwest continuation of the San Gabriel fault nd the area of its truncation of the San Juan-St. Francis (Morales) fault (as shown in figures 5E and 5F)? It should resumably have been displaced 225 km to the southeast along he San Andreas fault.

The final confirmation of any proposed scheme of fault move- nents depends on the determination that *all* rock units and eologic features along the faults are compatible with the pos- ulated sequence of events, and not compatible with any other cheme. This report, obviously, is based on only a small part of he geology within the large region under consideration. The onclusions of this report regarding the displacement histories of the San Gabriel and San Andreas faults are, however, in good accord with those given by other workers on the basis of entirely different evidence. The writer hopes that this is not just coinci- dence.

ACKNOWLEDGMENTS

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SYNTECTONIC BATHOLITHIC ROCKS IN SOUTH–CENTRAL SAN DIEGO COUNTY, CALIFORNIA, A PRELIMINARY REPORT[1]

By VICTORIA R. TODD[2]

ABSTRACT

Detoiled studies of rocks of the Peninsular Ronges botholith in the Cuyomoco Mountoins of south–centrol San Diego County disclosed thot oll plutonic rocks ore gneissic. The gneissic plutanic fobric resulted from synchronous deformotion ond recrystollizotian of rocks in o lorgely solid condition. The gneissic fobric extends beyond the Cuyomoco Mountains aver an area af at least 1200 km². The relotion between intrusive contocts ond plutonic foliotion suggests syntectonic intrusion ond continuing pastintrusive defarmation of the batholith ot elevoted temperoture. Pre–botholithic sedimentory rocks opparently underwent synkinemotic metomorphism during the some event in which the plutonic rocks were deformed and recrystollized. Thus, the region hos undergone o protrocted thermal and defarmotionol event which included syntectonic intrusion of the botholith, postintrusive defarmotian and synkinemotic metomorphism of wollrocks.

INTRODUCTION

This paper presents preliminary structural data for a part of the Peninsular Ranges batholith of southern California–the Cuyamaca Mountains in south–central San Diego County (figure 1). The Cuyamaca Mountains are part of a strip across the Peninsular Ranges, bounded by latitudes 32°45′N and 33°00′N, that is being mapped at present by the United States Geological Survey as part of an investigation of faulting in southern California. The area to be discussed coincides roughly with the Cuyamaca Peak 7½′ quadrangle.

Most geologic maps of San Diego County either show no structure in the plutonic rocks or depict it as primary–caused by flow of magma during consolidation of individual plutons (Everhart, 1951). Nomenclature used by Hurlbut (1935) and Larsen (1948) for plutonic rocks in the northern part of the southern California batholith has been applied by most workers to rocks in southern San Diego County. Specifically, inclusion–rich granitic rocks have been called Bonsall tonalite and inclusion–poor granitic rocks have been called Woodson Mountain granodiorite, with the result that mappable plutonic units have not been differentiated in many places.

Five plutonic and two metamorphic rock units have been mapped in the Cuyamaca Mountains during the present study and informal names have been assigned to avoid confusion with previous terminology (figure 2).[3] All of the plutonic units are foliated and most are gneissic; mylonite gneiss[4] occurs in intensely deformed zones. Study of rock textures indicates that foliation resulted from deformation and simultaneous recrystallization of a solid plutonic rock. Plotted foliations form a map pattern that is largely independent of individual pluton geometry indicating a regional deformation fabric (figures 2 and 3). The long dimensions of plutons commonly are parallel to the strike of the regional foliation suggesting that a stress field existed before and during emplacement. Local complex relations between intrusive contacts and plutonic foliation clearly indicate that some plutons were emplaced and deformed simultaneously. Significant deformation took place after plutons had solidified. Pre–batholithic metasedimentary rocks have a single synkinematic metamorphic fabric that is fully concordant with the deformational fabric of the plutonic rocks. Apparently the pre–batholithic sedimentary rocks and plutonic rocks were deformed and recrystallized during the same thermal and deformational event.

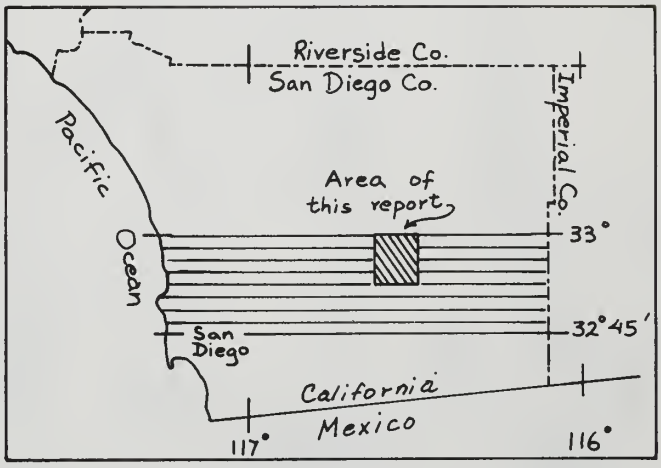

Figure 1. Index map of San Diego County showing location of Cuyamaca Mountains (diagonal lines) and east–west strip under study by U.S. Geological Survey (horizontal lines).

[1]Reviewed by D. Bukry, J. K.Crouch, D. M. Morton, and D. C. Ross.

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[3]Terminology applied to most plutonic and metamorphic rocks in this study follows the usage of Williams, and others (1954).

[4]A rock intermediate between mylonite and crystalline gneiss whose texture is the result of combined cataclastic deformation and recrystallization (Higgins, 1971).

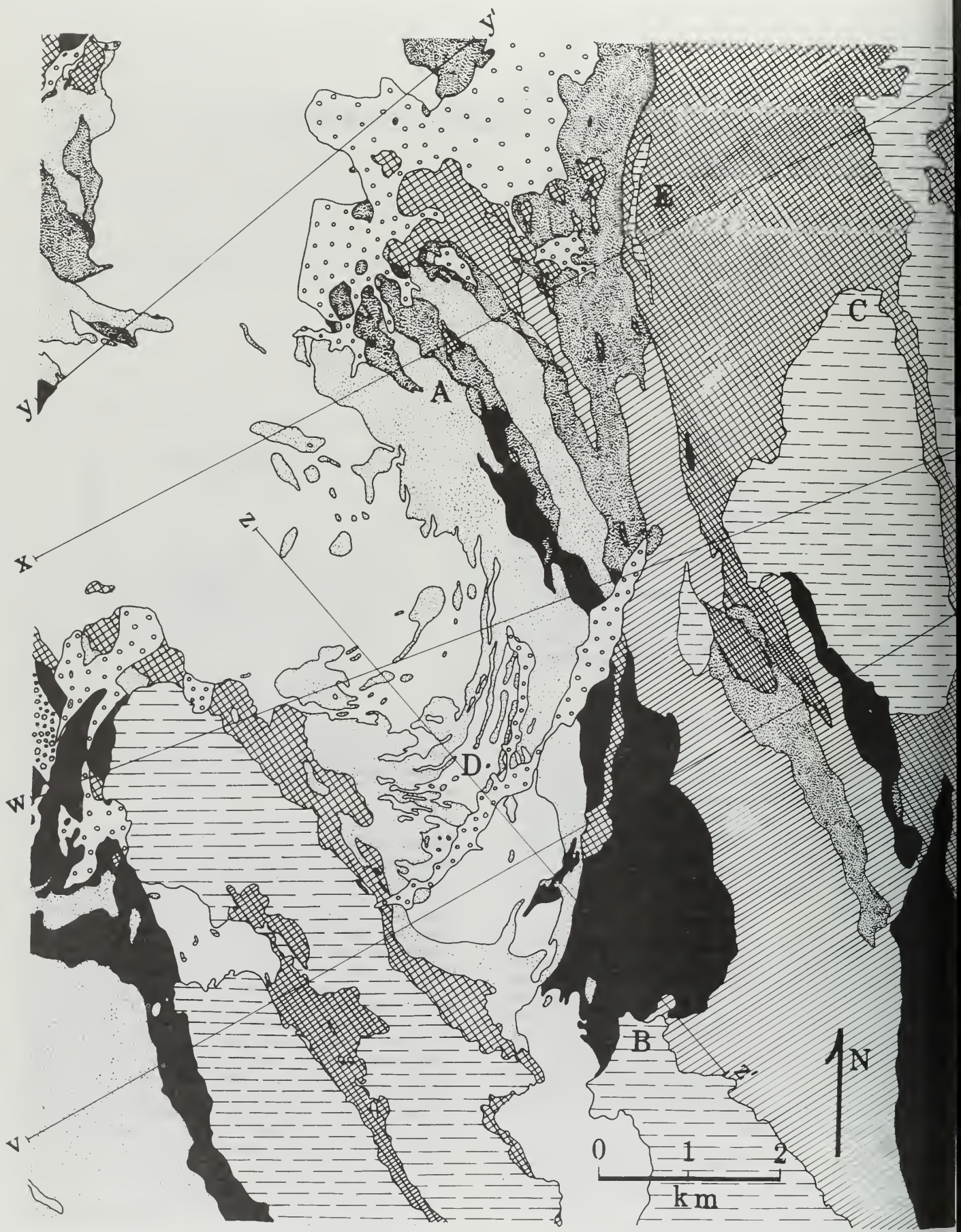
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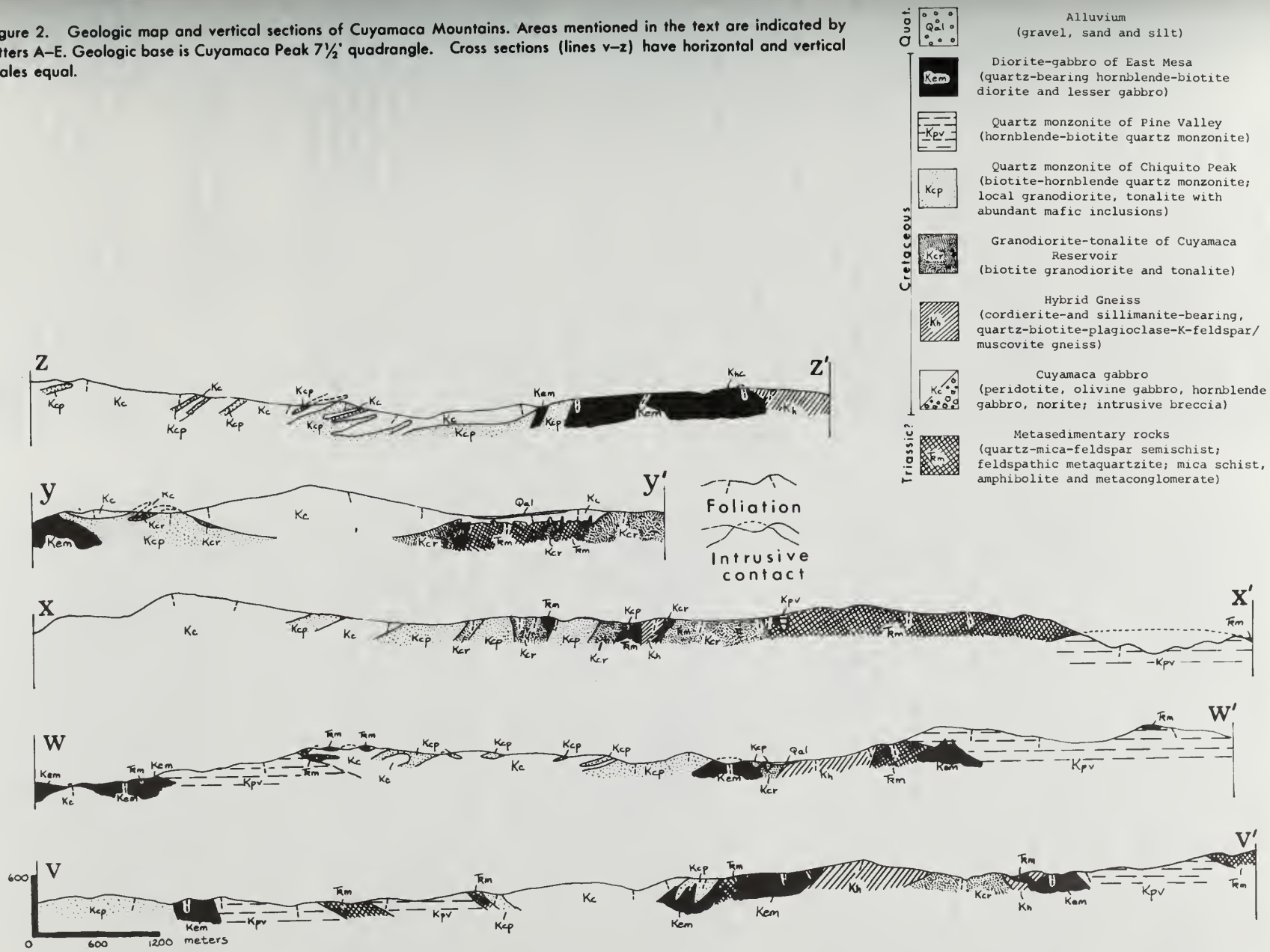
Figure 2. Geologic map and vertical sections of Cuyamaca Mountains. Areas mentioned in the text are indicated by letters A–E. Geologic base is Cuyamaca Peak 7½' quadrangle. Cross sections (lines v–z) have horizontal and vertical scales equal.



FOLIATION IN THE PLUTONIC ROCKS

Foliation in the plutonic rocks consists of a parallel arrangement of subplanar recrystallized mineral aggregates and relict phenocrysts (figure 4). Inclusions and schlieren[3] are oriented parallel to mineral foliation. Linear structures are present but commonly are difficult to measure. Steeply plunging lineation consists of elongate mineral grains and aggregates, and inclusions, aligned in the plane of foliation.

East of the map area, the plutonic gneisses grade into a broad zone of mylonite gneiss and schist (figure 5) forming a broad zone of rock that appears to have deformed strongly against a thick metasedimentary septum. Similar but narrower zones of mylonite gneiss occur at the margins of younger plutons where they intrude older plutons; these textures are also characteristic of dikes (figure 4). The fabric of the mylonite gneiss is fully concordant with, and gradational into, the gneissic fabric of the plutonic rocks and appears to be a more intense variant of the same deformation. Faults postdating the formation of mylonite gneiss are localized in the older zones of mylonite gneiss. The faults are marked by mylonite and gouge produced by brittle shear. Care must be taken to differentiate between mylonite gneiss and mylonite when evaluating suspected faults.

PETROGRAPHY

The textures of variably deformed plutonic rocks indicate that as deformation and recrystallization features increase, textural relics, such as euhedral grain boundaries and zones, and reaction relationships between mafic minerals, decrease (figure 5). Quartz, biotite and hornblende grains were crushed between large feldspar phenocrysts and recrystallized to aggregates of smaller grains. Interstitial aggregates in thin section have mosaic texture, indicating grain growth in the solid state (Kretz, 1966). Plagioclase phenocrysts were most resistant to recrystallization; even in strongly deformed rocks a few broken phenocrysts in a thin section retain euhedral oscillatory zones (figure 5).

In many thin sections, plagioclase grains bear multiple, thin, wedge–shaped twin lamellae that may be bent or broken (deformation twinning). As the number of such twinned grains increases, euhedrally zoned plagioclase relics decrease suggesting that deformation twinning aided in the obliteration of euhedral zoning. Concurrently, plagioclase phenocrysts underwent marginal embayment by surrounding minerals and recrystallization to fine–grained aggregates (figure 5).

STRUCTURE OF CUYAMACA MOUNTAINS

The north–northwest structural grain imparted by linear shapes of younger plutons is an obvious feature of the map of figure 2. Because the map area is small, the structural grain of the older plutons is not as obvious. However, the same regional structural grain is shared by all granitic plutons and probably by the large gabbro masses in south–central San Diego County. South and southwest of the Cuyamaca Mountains the structural grain swings gradually from north–northwest to northwest, east–west, and east–northeast so that the overall structure of the batholith is complex. Local deviations from the structural grain occur (D on the geologic map, figure 2). Preferred orientation of plutons in part could be inherited from a pre–existing structural grain, but the relations between pluton foliation and intrusive contacts indicate that much intrusion was syntectonic and that deformation continued after intrusion.

Foliation in individual plutons commonly is oriented parallel to walls (Stonewall Peak area, A on geologic map, figures 2 and 3), but in many places the attitude of foliation passes without deviation from one pluton into another (East Mesa, Rattlesnake Valley, B and C on geologic map figures 2 and 3). The regional foliation pattern is therefore largely independent of interpluton contacts and cannot be due alone to protoclastic flow within single plutons.

The intrusive contact between quartz monzonite of Chiquito Peak and Cuyamaca gabbro, well exposed on the east flank of Cuyamaca Peak (Arrowmakers Ridge, D on geologic map, figures 2 and 3), is probably syntectonic. The strike of foliation in both units follows intrusive contacts suggesting that the deformation which produced the foliation was guided by the complex shape of this contact and that gabbro and quartz monzonite were both mobile (yet solid) at this time. The dip of foliation is much steeper than that of the intrusive contacts (figure 2B), a puzzling relationship that is observed in many places where the strikes of foliation and intrusive contacts are concordant.

The fabric of primary plutonic structures such as dikes records both synintrusive and postintrusive deformation. On a small map scale, dikes generally were emplaced parallel or sub–parallel to regional foliation and dike foliation follows regional foliation (E on geologic map, figure 2). Smaller dikes, seen at a single outcrop, may crosscut host rock (regional) foliation, but are folded isoclinally with their axial planes parallel to foliation (figure 4). Dike foliation itself is always concordant with that of the host rock and passes through the hinges of folds. The preservation of such primary structures indicates that deformation was the result of small incremental movements distributed throughout a large volume of rock.

The plutonic fabric is concordant with that of the metamorphosed sedimentary rocks which form roof pendants and septa in the batholith (figure 2). The metasedimentary rocks locally contain well–preserved sedimentary structures (graded bedding, cross–bedding) which were overprinted by a single synkinematic metamorphic fabric in which metamorphic foliation parallels relict bedding. The concordance of the deformational fabric in the plutonic rocks with the synkinematic metamorphic fabric in the sedimentary rocks, and the absence of an earlier metamorphic fabric in the sedimentary rocks indicate that, in this part of the batholith, plutons and their metasedimentary wallrocks were deformed and recrystallized simultaneously.

MODE OF EMPLACEMENT OF PLUTONIC UNITS

An indirect line of evidence suggesting a close relationship between intrusive and deformational processes is the change in plutonic style during emplacement of the batholith. In general early plutons have broad marginal zones of assimilation–contamination reactions with older rocks while younger plutons have discrete margins or only narrow reaction zones. Each unit has a characteristic intrusive relation to older rocks yet all show the effects of the same regional stress. This suggests interplay between intrusive and deformational events rather than reordering of an already existing batholith during a later, unrelated tectonic event.

[3]Schlieren are streaky, wispy or nonstratiform layers that differ from the host rock in composition, texture, or both (Compton, 1962).

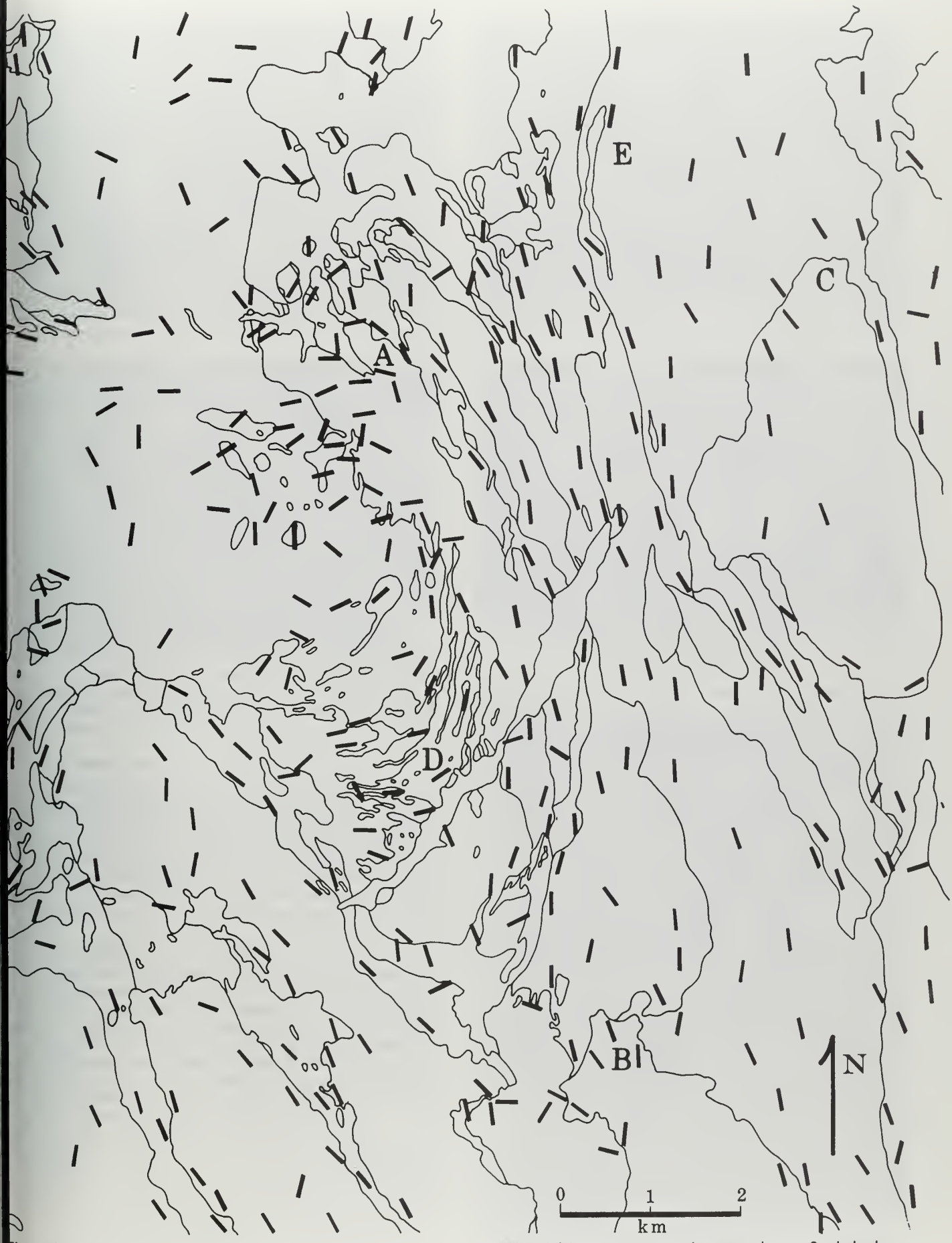


Figure 3. Structure map of Cuyamaca Mountains. Faults have been omitted. Solid lines are contacts between plutons. Symbols show generalized strike and dip of foliation. Dips are steep, 70–90°. Areas mentioned in the text are indicated by letters A–E.

Cuyamaca Gabbro

Gabbro in south–central San Diego County underlies a number of solitary peaks which appear to be isolated plutons. Everhart (1951) named the gabbro bodies of the Cuyamaca Mountains the Cuyamaca gabbro. He considered the Cuyamaca gabbro to be younger than the quartz plutonites of the Stonewall Peak area (A on geologic map, figure 2), but the present study has found that the granodiorite–tonalite of Cuyamaca Reservoir consistently has a finer grained margin, 1–2 m wide, against the gabbro, intrudes the gabbro in dikes, and locally contains gabbro inclusions; the granodiorite–tonalite is demonstrably the oldest of the quartz plutonites (figure 2, map legend). When younger units are removed, the pattern made by metasedimentary inclusions (the wallrocks of the gabbro) in south–central San Diego County suggests that gabbro bodies were elongate parallel to the regional structural grain prior to intrusion of the quartz plutonites .

The age relations between gabbro intrusion and regional deformation remain unsolved. Much if not all of the gabbro of the Cuyamaca Mountains has a strong mineral foliation that in many places is parallel or sub–parallel to apparent primary igneous layering (including hornblendic schlieren). Rock textures indicate only mild strain effects and moderate recrystallization––probably the textures described by previous workers as protoclastic (Nishimori, 1976). In many places gabbro foliation is concordant with regional foliation (northwest map area) but elsewhere it diverges or appears complex. Foliation is either absent or difficult to detect in nearly equigranular gabbro. Local divergence of gabbro foliation from the regional fabric may represent a relict primary igneous fabric, local secondary deformation, or complex response of the large volume of mafic rock to stress.

Granodiorite–Tonalite of Cuyamaca Reservoir

The granodiorite–tonalite of Cuyamaca Reservoir intruded metasedimentary rock along steeply dipping bedding (foliation) planes and assimilated large volumes of it to form broad (2–3 km) zones of hybrid, cordierite– and sillimanite (fibrolite)–bearing quartz–biotite–plagioclase–K–feldspar (or muscovite) gneiss which have complex, interfingering contacts with both units. The hybrid gneiss is homogeneous on a small scale and contains relatively little migmatite. Evidence of igneous parentage of the gneiss includes local phenocrysts of K–feldspar containing zonary euhedral plagioclase inclusions and euhedrally zoned plagioclase phenocrysts (figure 5). Such grains are not seen as relict detrital feldspars in the metasedimentary rocks, nor are they porphyroblasts. The abundant quartz and mica, sedimentary inclusions, and presence of cordierite and sillimanite attest to the rock's partly sedimentary origin. The granodiorite–tonalite and hybrid units are the most intensely deformed on the quartz plutonites; hybrid gneiss is mylonitic near contacts with metasedimentary rock. Tightly appressed isoclinal passive flow folds with axial planes parallel to foliation occur in both units. The granodiorite–tonalite and hybrid units are so deformed that they were previously thought to pre–date the mid–Cretaceous batholith; in many places they are similar–appearing and were mapped together as "Mixed Rock" (Everhart, 1951).

Quartz Monzonite of Chiquito Peak

The quartz monzonite of Chiquito Peak forms both large, steep–walled, north–northwest–trending dikes and more gently dipping interconnected wedge–shaped bodies in older rocks (Stonewall Peak area, A; Arrowmakers Ridge, D on geologic map). Structure sections across the Cuyamaca Mountains suggest that the gabbro is in part rootless–resting upon a substratum of younger units, primarily quartz monzonite and granodiorite–tonalite of Cuyamaca Reservoir (y–y′ and z–z′, figure 2B). The geologic map does not show topographic contours, but dips of gabbro–quartz monzonite contacts follow contours closely on both east and west sides of the Cuyamaca Mountains and are therefore gently dipping. The change in dip, along strike, of the quartz monzonite's contacts with older units may reflect the quartz monzonite's response to inherent structural properties of the rocks it intruded, e.g., to steep foliation in the granodiorite–tonalite of Cuyamaca Reservoir. Although no shallow primary structure was observed in the gabbro, there may be cryptic shallow structure or compositional zonation that guided the intrusion of the quartz monzonite. Another possibility is that the quartz monzonite was emplaced in tongue–like bodies which have both steep–walled, rooted, and gently dipping, sheeted forms. The true form in three dimensions of this widespread unit cannot be determined from these map relations alone.

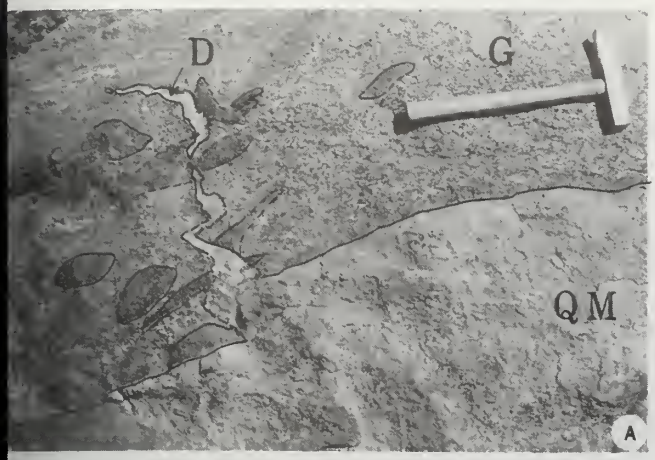
Typically, marginal parts of the quartz monzonite plutons are contaminated to granodiorite and tonalite and choked with flattened mafic inclusions, many of which are gabbroic. The inclusions are disk–shaped and aligned parallel to steep foliation, and on outcrop surfaces that are nearly perpendicular to foliation they appear as swarms of elongate spindles. Where the unit intrudes gabbro, contaminated zones are broad, e.g., in Tule Springs 7½′ quadrangle, west of the map area, where Chiquito Peak–type quartz monzonite can be traced westward into granodiorite and tonalite . The adjacent gabbro bodies, here the gabbro of Cuyamaca Peak, are bordered by broad zones of deformed intrusive breccia consisting of partly assimilated gabbro blocks in contaminated quartz monzonite (figure 2). Locally these two marginal zones are gradational. Similarly, where the quartz monzonite has intruded granodiorite–tonalite of Cuyamaca Reservoir, linear bands of contaminated quartz monzonite (chiefly intimately interlayered quartz monzonite and fine–grained tonalite) occur between the two units. Quartz monzonite of Chiquito Peak has also assimilated metasedimentary rock locally to form lenses of distinctive hybrid hornblende–plagioclase rock of heterogeneous composition and texture.

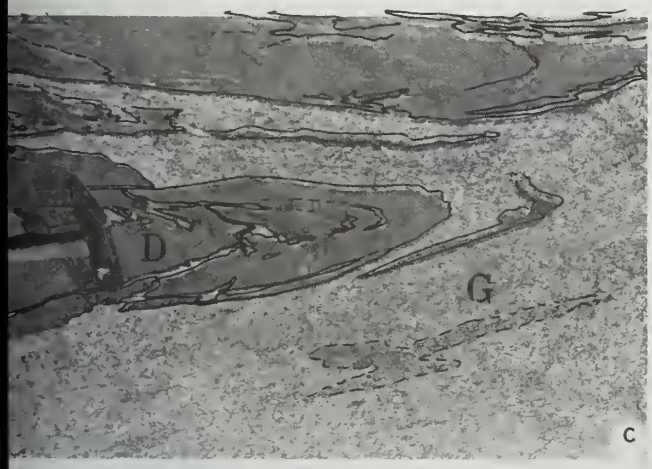
Quartz Monzonite of Pine Valley

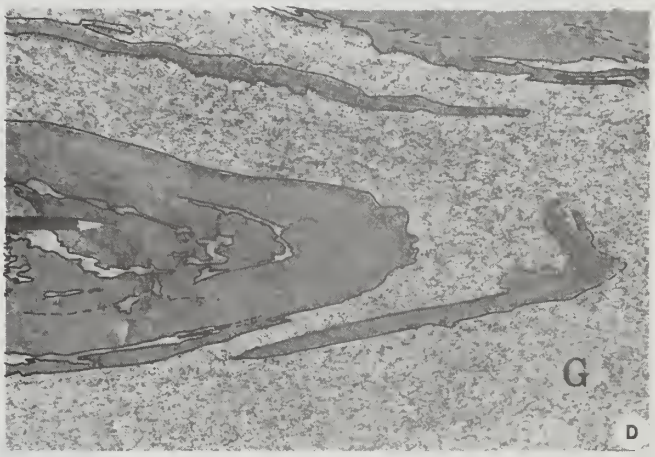
In contrast to the older quartz plutonites, the quartz monzonite of Pine Valley forms discrete, homogeneous plutons which intrude the surrounding rocks cleanly, contain few marginal inclusions and are uncontaminated. The plutons are linear, elongated north–northwest, and have lobate terminations; generally they are steep–walled but locally send more gently dipping tongues into older rocks (figure 2B). The unit is as strongly deformed and recrystallized as the older quartz monzonite.

Diorite–Gabbro of East Mesa

Elongate plutons of quartz–bearing diorite comprise the youngest plutonic unit of the Cuyamaca Mountains (figure 2). The plutons are strongly deformed internally—zones of mylonitic rock occur at the margins while their cores may be mylonitic or igneous–appearing. Textures vary greatly, but modal minerals of the unit are consistently, in order of decreasing abundance, plagioclase, subequal hornblende and biotite, quartz, minor K–feldspar and accessory sphene; rocks with pyroxene relict cores in hornblende and biotite grade into the less abundant gabbro of the unit. The plutons typically have intruded along contacts between two or more older units and their shapes reflect the geometry of these contacts. For example, the linear north–northwest–trending East Mesa pluton (B on geologic map, figure 2) follows contacts of hybrid gneiss, granodiorite–tonalite of Cuyamaca Reservoir and quartz monzonite of Chiquito Peak, tapering northward from a bulbous termination against quartz







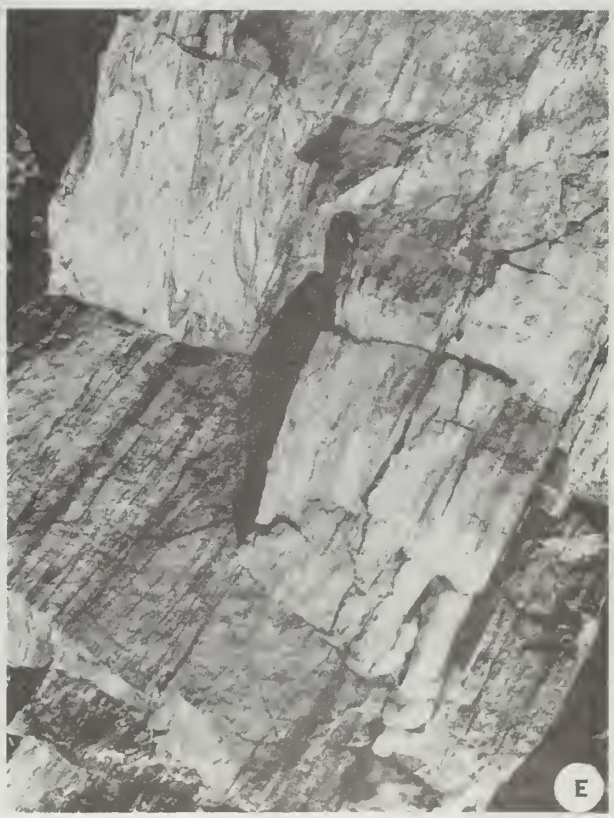
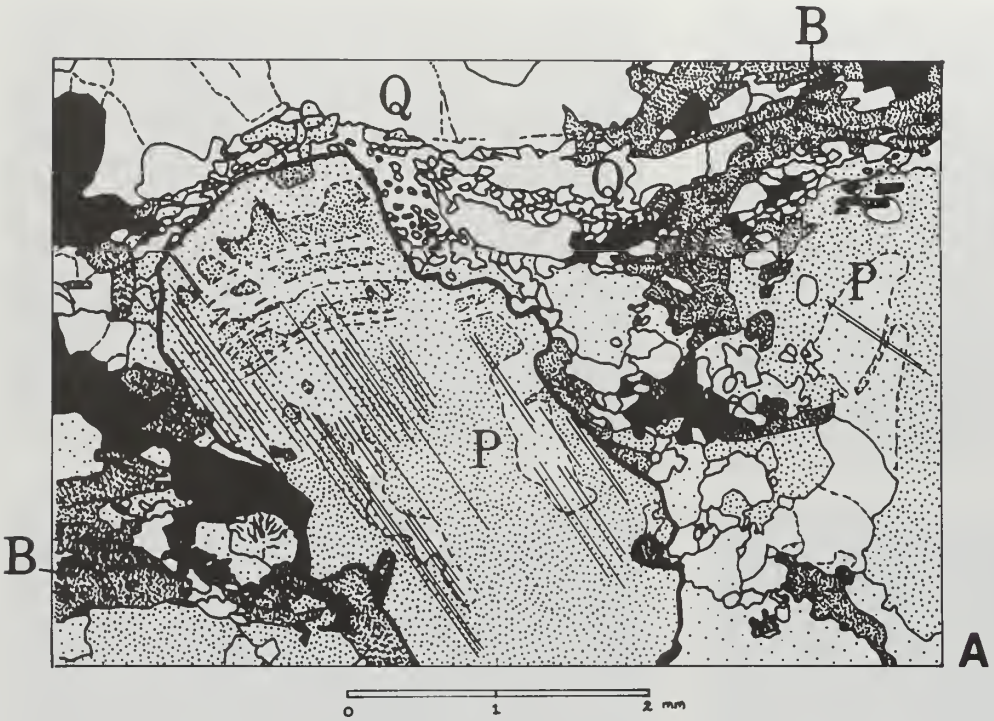


Figure 4. Textural variations in plutanic racks af the Cuyamoco Mauntains. *A,* Contoct between chilled quartz manzonite of Chiquito Peok (QM) and granadiarite–tanalite af Cuyamaca Reservair (G); hammer handle is parallel ta foliation of mineral groins ond oligned mafic inclusians in gronadiarite–tanalite, ta faliatian in quartz manzanite, and ta axial planes af falds in quortz manzanite dikelet (D). *B,* Clase view af fald in quartz manzanite dikelet, pencil parallel ta foliation. *C,* Falded synplutonic fine–grained dike af diarite–gabbra of Eost Meso (D) in gronodiarite–tanolite of Cuyamaco Reservair (G); hammer hondle parallel ta faliation in bath rocks ond to axiol plane of synform. *D,* Close view of some synplutonic dike, pencil por allel to oxial planes af falds in chilled veinlets af gronadiorite–tonolite (G). *E,* Port of mylonitic dike af quortz manzanite of Pine Volley; pencil (right–hand side af photogroph) is parallel ta trace af faliatian which dips steeply ta left; nate isaclinol falds, upper left; dark ond light banding is coused by segregation af camminuted and recrystallized minerols.



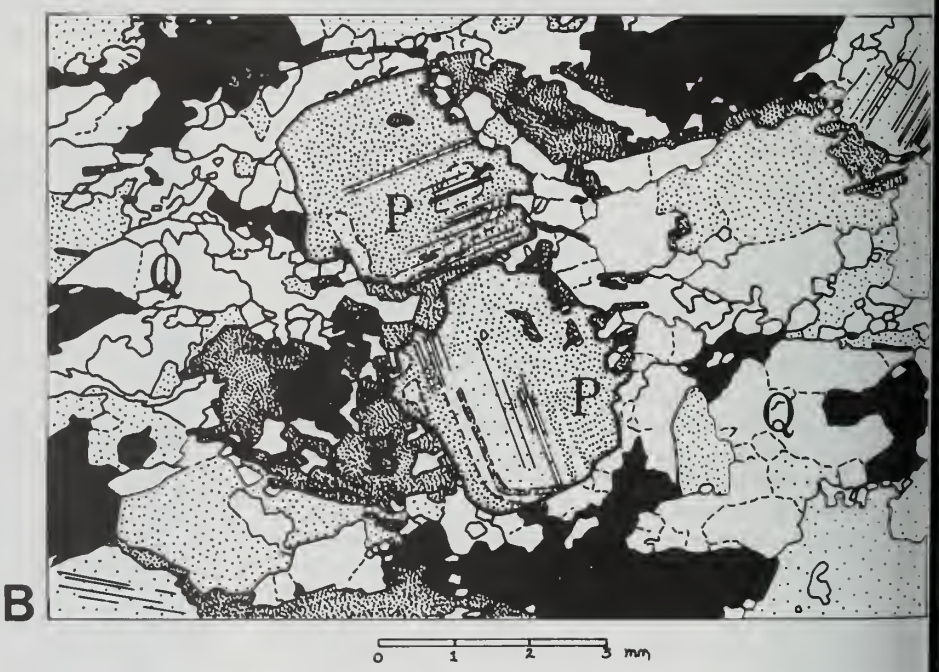


Figure 5. Photomicragraphs showing textural variatians in plutonic racks. *A,* Large relict plagiaclase phenacryst (P) with oscillatory zones and deformatian twinning; granulated and recrystallized quartz (Q) and biatite (B) abave, between twa large grains. Quartz monzonite of Chiquita Peak. Crassed Nicals. *B,* Twa relict plagiaclase phenacrysts (large dark grains) in mylonitic hybrid gneiss. Oscillatory zones transected by grain margin (lower grain). Quartz and biotite are largely recrystallized.

monzonite of Pine Valley to a feather–edge of large dikes in the Stonewall Peak area (A on geologic map).

The chilled margins of the diorite–gabbro bodies have been synplutonically intruded by their granitic host rocks and deformed such that intrusive age relations often are reversed. Some "veins" of granitic rock in the margins of diorite bodies are actually deformed and partly assimilated inclusions of granitic rock, while chilled diorite margins and dikes have been intruded by host rock, infolded and pulled–apart so that they resemble inclusions (figure 4). Only careful study of many outcrops revealed the true age relation. In general, chilling of younger rocks against older was found to be the most reliable criterion of intrusive sequence because of post–emplacement deformation and recrystallization.

A widespread system of diabase dikes locally intrudes the diorite–gabbro unit but is probably closely related to it in age because in some places diabase dikes grade into diorite plutons. The dikes contain deformed bodies of aplite that locally are partially continuous with, and a finer grained phase of, surrounding quartz plutonites; more often, these aplite bodies are confined strictly to the diabase dikes. Both diabase dikes and aplite bodies share the regional fabric of their host rocks. Many diabase dikes have been intruded and deformed by host rock and resemble inclusions—as described above.

Although the total area of outcrop of the diorite–gabbro of Japatul Mesa and associated mafic dikes is much less than that of older plutons, these mafic rocks form an ubiquitous network and may represent only the upper part of a much larger volume of rock at depth.

DISCUSSION

Mapping and petrographic study of rocks of the Peninsular Ranges batholith in the Cuyamaca Mountains revealed that all of the plutonic rocks are foliated and most are gneissic as a result of synchronous deformation and recrystallization in a largely solid condition. The resulting fabric is present throughout a minimum area of 1200 km^2.

The relations between plutonic foliation and contacts in the Cuyamaca Mountains suggest that the fabric resulted from synorogenic intrusion and continued deformation at elevated temperatures rather than from simple protoclastic flow within plutons. There was a change in intrusive style from early plutons with broad chemical and mechanical reaction zones to late plutons with restricted reaction zones. The entire suite of plutonic rocks displays the same regional structural control on pluton shapes and foliation suggesting syntectonic intrusion rather than overprint of tectonite fabric. If the batholith had been emplaced, consolidated, and at a later time, reheated and deformed, more disordering or overprinting of intrusive contacts and plutonic structures would be expected. The deformational fabric can be traced from plutonic rocks into synkinematically metamorphosed sedimentary rocks without deviation, making it unlikely that metamorphism of the sedimentary rocks was widely separated in time from intrusion in this part of the batholith.

The above data imply a long thermal and deformational event which included batholithic emplacement and metamorphism of sedimentary wallrocks. Schwarcz (1969) found that metamorphism of sedimentary rocks preceded intrusion of the batholith in Riverside County to the north. Silver and others (1969) recorded a complex batholithic history in northwestern Baja California which includes both late–syntectonic intrusion (115 ± 3 m.y., U–Pb zircon ages) and post–tectonic intrusion (continuing until slightly less than 100 m.y. ago). Thermal and deformational events thus appear to have fluctuated across the batholith. Further studies should clarify the time of intrusion, metamorphism and deformation.

K–Ar ages of rocks of the Peninsular Ranges batholith are generally younger than U–Pb zircon ages and have been interpreted as recording time of cooling rather than time of emplacement (Krummenacher and others, 1975). Recrystallized hornblende and biotite from plutonic rocks of the Cuyamaca Mountains yielded K–Ar ages ranging from 71–108 ± 3 m.y. (W. Hoggatt and V. R. Todd, unpublished data, 1976). These dates may be considered metamorphic ages in view of the above data.

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I wish to thank Wendy Hoggatt for assistance in the field and for discussion of many ideas presented here. The field study would not have been possible without the cooperation of rangers of the Cleveland National Forest (United States Forest Service) and the Cuyamaca Rancho State Park (California Department of Parks and Recreation) who gave advice and practical aid whenever asked. Landowners who have generously allowed access to their property are too numerous to name—but their goodwill is nonetheless appreciated. D. Bukry and J. K. Crouch read a rough draft of the manuscript and made many suggestions which improved it. The writer thanks D. M. Morton and D. C. Ross for critical reviews of the manuscript.

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ETROGRAPHY AND METAMORPHISM OF THE MORENA RESERVOIR OOF PENDANT, SOUTHERN CALIFORNIA[1]

RICHARD G. BERGGREEN[2] and MICHAEL J. WALAWENDER[2]

ABSTRACT

Morena Reservair roof pendant consists af regianally metamarphased ortzo-feldspothic units that are defarmed inta a tightly folded antifarm. ese units ronge from cardierite-biotite schist with relict sedimentary struc- es ta diapside gneiss and metacanglamerate. Preserved sedimentary fea- es indicote depositian in a shallaw-water enviranment. Metamorphism and farmatian occurred before plutanic elements af the Peninsular Ranges tholith were emplaced. Metamarphic parageneses suggest Abukuma-type tamarphism and bracket $P_f = P_s$ — T canditians at 2.0–2.5 kb and 600°C.

INTRODUCTION

The prebatholithic rocks of the Peninsular Ranges consist of series of regionally metamorphosed volcanic-volcaniclastic and dimentary units that were intruded during late Jurassic to iddle Cretaceous time by the Peninsular Ranges batholith. nroofing of the batholith has left only isolated roof pendants the prebatholithic units. In San Diego County, the volcanic units (Santiago Peak Volcanics, Larsen, 1948) occur along the western margin of the batholith, the central zone is characterized by the presence of pelitic units, and the eastern zone by pelitic units with significant carbonate content. The metasedimentary rocks of the central zone (Julian Schist, Donnelly, 1934) have been prospected and mined for gold since the 1880's, but extensive studies on the structural, petrographic, and metamorphic character of these units have not been undertaken. Gastil and others (1975) and Berggreen (1976) have summarized the previous work on the metamorphic rocks of southern California. This paper reports on recent work in one roof pendant and its implications regarding the prebatholithic history of southern California.

The Morena Reservoir roof pendant (figure 1) is surrounded by plutonic rocks typical of the Peninsular Ranges which include the Los Pinos Gabbro (Walawender, 1975, 1976), the Bonsall Tonalite, and the Woodson Mountain Granodiorite (Everhart, 1951). The roof pendant is an elongate septum, approximately 8 km in its longest dimension and averaging about 2 km in width. The long dimension of the body is parallel to the north-northwest structural grain of the Peninsular Ranges.

eviewed by Victoria Todd and Gordon Gastil.

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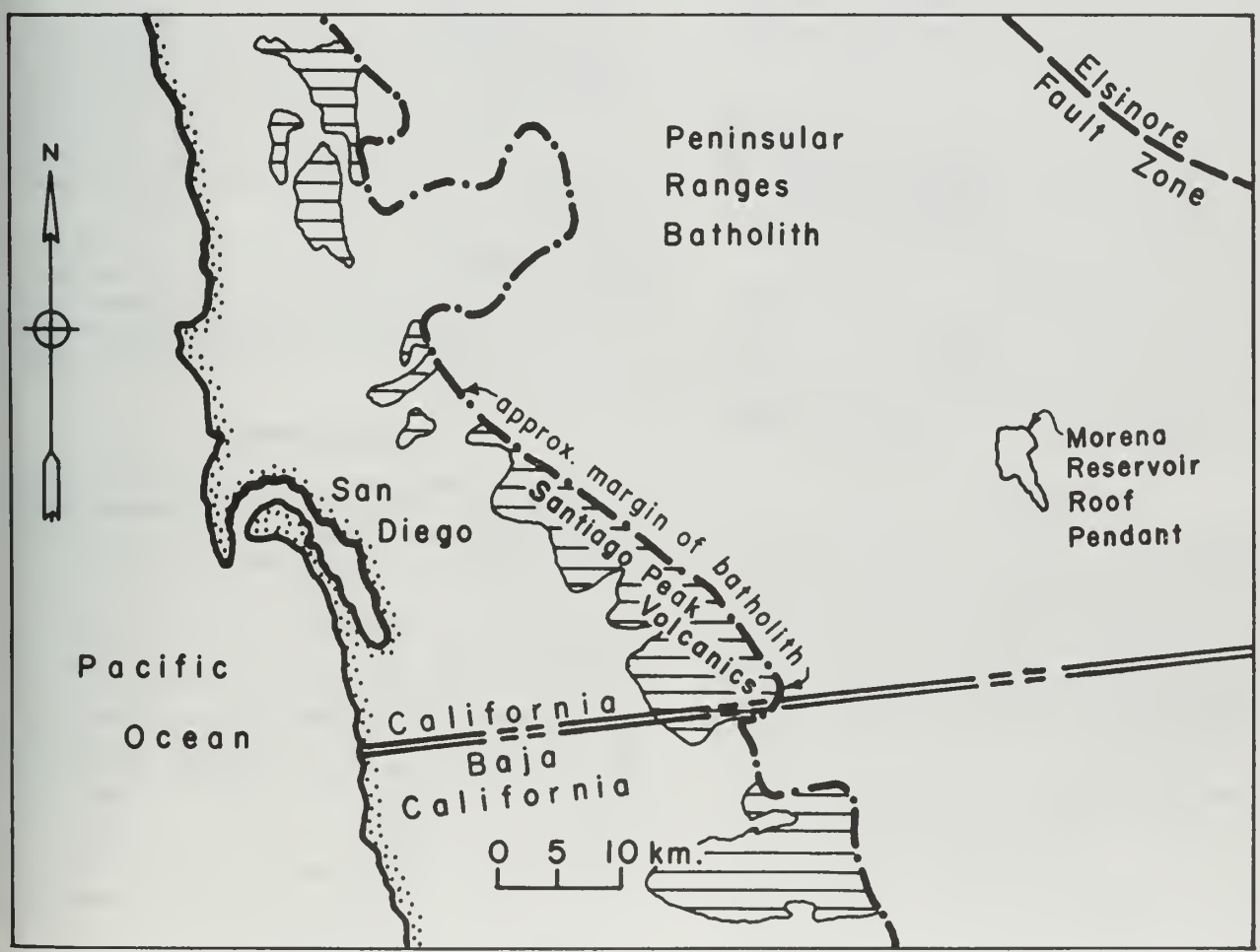

Figure 1. Locatian map af the Morena Reservair raof pendant.

The metasedimentary rocks are well exposed along steep slopes at 3,200 to 4,000 foot elevations and occur locally to elevations over 4,200 feet. These topographic highs are on what is recognized by Gastil (1961) as a pre-Pliocene erosion surface. North of the roof pendant, this erosion surface is well preserved in Corte Madera Valley and Long Valley which have an average elevation of about 4,000 feet. The excellent exposures are a result of headward erosion onto this surface from Morena Valley on the south. Parts of the roof pendant are exposed in the erosion scarp between the two levels although the best exposures are in the narrow canyons incised into the higher level. The northern portion of the roof pendant in Corte Madera Valley and Long Valley is poorly exposed.

STRUCTURE

The metasedimentary rocks are tightly folded into a northwest-plunging, asymmetrical antiform whose axis is parallel to the long dimension of the roof pendant (figure 2). The fold is characterized by a steeply dipping western limb, a series of complex folds near the axis, and a steeply dipping to vertical eastern limb. The present erosion surface appears to be below the turnover of the fold although it may be preserved to the northwest where the metasedimentary rocks are poorly exposed. Numerous smaller folds in the axial plane region suggest that the larger structure plunges to the northwest at angles of 5 to 15°. To the southeast, the axial region contains smaller bodies of the granitic rocks and is characterized by an intimate mixture of igneous and metamorphic rocks.

On the western side of the roof pendant, there is a prominent lobe of metasedimentary rocks. This lobe has the same rock types as the main portion of the pendant but the strike of the foliation is almost perpendicular to the structural trend found in the main body of the metasedimentary rocks to the east. Moreover, dips of the foliation appear to be shallower on the northern end of the lobe than on the southern end. The antiformal structure that characterizes the main segment of the roof pendant is not present in the western lobe. These data suggest that the western lobe is composed of the same sequence of metasedimentary rocks but that it is in structural contact with the main portion of the roof pendant. The specific nature of this contact is undetermined.

A system of northwest trending faults of undetermined slip cut both the metasedimentary rocks and the surrounding igneous units. Merriam (1958) described a series of faults with the same trend in the Santa Ysabel quadrangle 40 km to the northwest. The Hauser Canyon lineament, about 5 km to the southwest of the Morena Reservoir roof pendant, is also parallel to the faults in the Morena Reservoir area.

PETROGRAPHY

The metasedimentary rocks exposed within the Morena Reservoir roof pendant consist mainly of quartzo-feldspathic schist and gneiss. The schistose rocks are the most abundant and are characterized by biotite and cordierite. The gneiss and rare metaconglomerate form a small fraction and contain up to 33 percent calc-silicate minerals. Biotite is rare and cordierite absent in the gneiss, whereas the schistose units are lacking the calc-silicate minerals. Modal compositions of typical samples are given in table 1.

Relict sedimentary bedding is preserved throughout the section and is essentially parallel to the foliation. It ranges from a few millimeters to more than one meter in thickness and is generally more massive in the quartz-rich varieties. Ripple marks, cross stratification, and graded beds are preserved locally but much of their finer detail has been destroyed during metamorphism. Minute inclusions of graphite outline cores in many of the quartz and feldspar grains of the schistose rocks and indicate that the original clastic particles were subangular to well rounded and ranged in size from medium to fine sand. Quartzo-feldspathic pebbles occur throughout the section but are rare in the schistose units.

The cordierite-biotite schist is generally well foliated and contains varying amounts of andalusite, sillimanite, muscovite, and plagioclase (An_{30-45}). Although the two alumino-silicates are often found in the same samples, they exhibit different modes of occurrence. The sillimanite is present as felted mats of fibrolite subparallel to associated biotite grains. The andalusite occurs as large poikiloblasts up to several millimeters across that enclose optically continuous patches of potassium feldspar. Andalusite is more abundant in the northwestern end of the roof pendant whereas sillimanite predominates in the southeastern end. Muscovite occurs as a finer-grained phase intimately intergrown with both alumino-silicates and as rare porphyroblasts oriented at high angles to the foliation. The latter does not appear to exhibit a reaction texture with either sillimanite or andalusite. Almandine garnet was found only in a small septum at the northeastern end of the roof pendant. Staurolite was not observed in any of the samples.

The gneiss and the matrix of the metaconglomerate units have granoblastic textures and contain diopside plus varying amounts of actinolite, clinozoisite, and plagioclase (An_{50-70}) with minor calcite and wollastonite. These units are best exposed near the center of the roof pendant along the axis of the antiform but also occur as isolated outcrops within the schistose units. Metaconglomerate lenses occur locally up to 3 to 5 m in thickness and are generally less than 20 m in length. The clasts in the metaconglomerate are composed of quartz. They are well rounded, coarsely crystalline, and range from granule to cobble size. They vary from spherical to ellipsoidal but each lens of the metaconglomerate appears to have a restricted range in clast shape. The finer-grained gneissic units exposed in the axial region are often complexly folded with amplitudes ranging from less than a centimeter up to several meters. The fold axes are parallel to that of the major antiformal structure and plunge northwest at angles of 5–15°.

Diopside and amphibole are common throughout the gneissic units. The amphibole has low second-order interference colors, high 2V (75–85°), weak pleochroism in shades of pale green, and a maximum extinction angle of 16–18° which suggest an actinolitic composition. The diopside usually occurs as a corona around the amphibole.

METAMORPHISM

The various mineral associations and textural relationships in the rocks of the Morena Reservoir roof pendant suggest a series of metamorphic reactions. These reactions have been studied experimentally by other investigators and can be used to bracket P — T conditions.

The coronas of diopside around actinolite in the gneissic units indicate the reaction

(1) actinolite + calcite + quartz = diopside + CO_2 + H_2O.

The depletion of available carbonate may be responsible for the equilibrium preserved between these minerals. A similar reaction involving a tremolitic rather than actinolitic amphibole has been studied by several investigators (Turner, 1967; Metz and Trommsdorf, 1968; Winkler, 1974). Although analyses of the Fe/Mg ratio of the amphibole in the Morena Reservoir roof pendant are not available, it is probable that this phase contains some iron in solid solution. This component would tend to shift the univariant curve for the reaction to lower temperatures so that reaction 1 (figure 3; after Winkler, 1974) approximates an upper limit for metamorphic conditions.

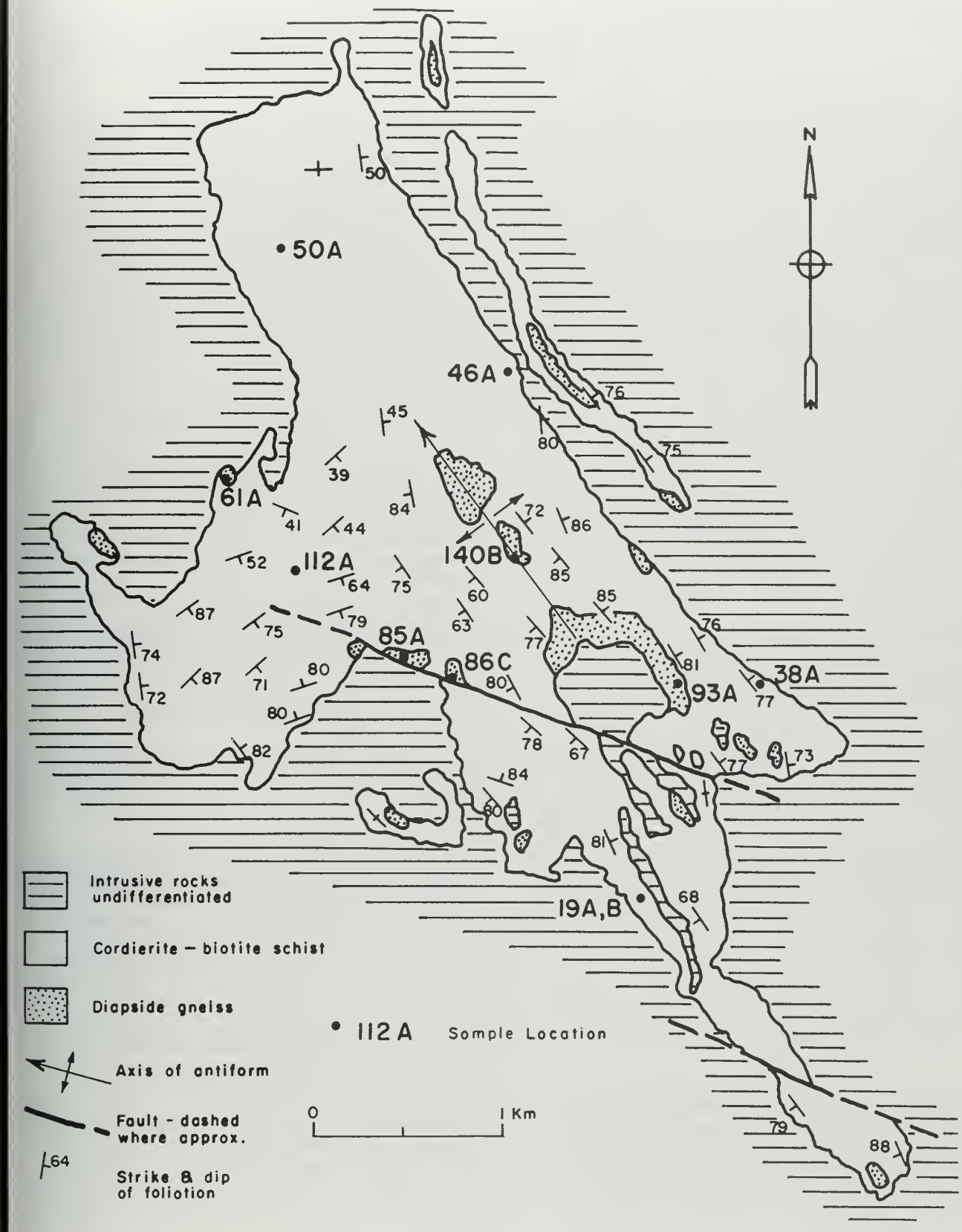

Figure 2. Geologic map of the Morena Reservoir roof pendant (after Berggreen, 1976).

Table 1. Modal composition of representative samples from the Moreno Reservoir roof pendant. See figure 2 for sample locations.

	Cordierite-biotite schist					*Diopside-actinolite gneiss*					
	19A	*38A*	*19B*	*46A*	*50A*	*112A*	*140B*	*85A*	*86C*	*93a*	*61a*
Quartz	62	52	23	32	22	36	50	62	53	55	40
Biotite	14	16	5	19	16	15	2				
Cordierite	2	16	24	8	10	13					
Plagioclase	20	10	15	26	17	17	19	3	19	24	40
K-Feldspar	Tr	4	23	6	18	7	13	Tr	1		
Muscovite		1	3	1	2	5					
Andalusite				3	5	1					
Sillimanite		Tr	5	4	2	Tr					
Diopside							3	20	23	10	10
Actinolitic Amphibole						6			10		
Clinozoisite							1	2	1	1	5
Sphene							Tr	Tr	1	Tr	2
Wollastonite								1			
Calcite								1			
Apatite	Tr	Tr	Tr	Tr		Tr	Tr	Tr	Tr		2
Grossularite								9			
Graphite	1	Tr	1	1	3	4	2	2	2	Tr	1
Tourmaline	1	Tr			Tr						
Zircon	Tr			Tr							
Magnetite					5	2	3				

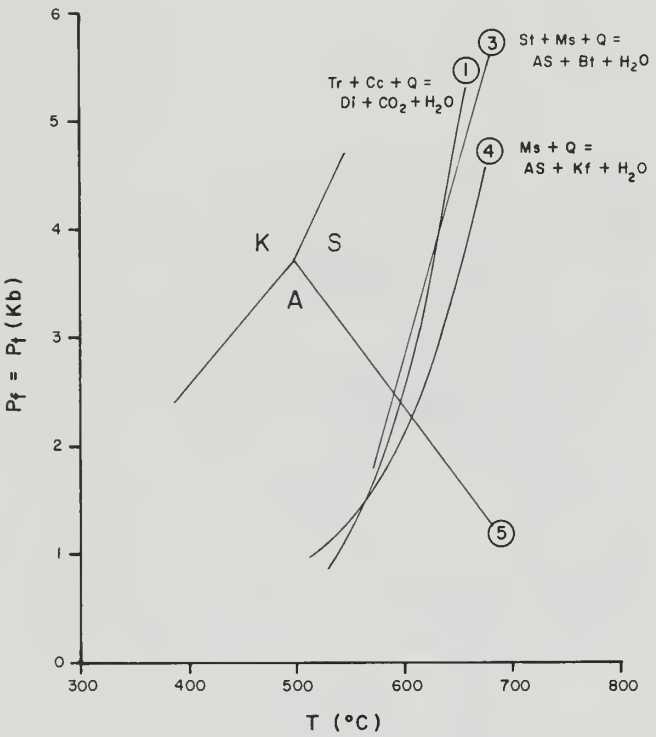

Figure 3. Petrogenetic grid. Numbers keyed to reactions discussed in text.

The occurrence of andalusite and sillimanite in two different associations in the Morena Reservoir roof pendant suggests that they are not derived via polymorphic inversion but by two different metamorphic reactions. Schwarcz (1969) recognized the association of sillimanite and biotite in the prebatholithic rocks of the French Valley Formation in the northern Peninsular Ranges and suggested that the association may result from the reaction

(2) biotite$_1$ + quartz = biotite$_2$ + olumina-silicate + H_2O

where biotite$_1$ is more aluminum-rich than biotite$_2$. The association of the alumino-silicates and biotite plus the absence of staurolite, however, can both be accounted for by the reaction

(3) stourolite + muscovite + quartz = olumina-silicote + biotite + H_2O.

The P_{H_2O} — T curve for this reaction (Hoschek, 1969) is given in figure 3.

The relationship between andalusite and potassium feldspar (± muscovite and quartz) appears to indicate the reaction (Evans, 1965).

(4) muscavite + quartz = andalusite + patassium feldspar + H_2O.

The predominance of sillimanite over andalusite in the southwestern (more deeply eroded) end of the plunging antiform and the systematic increase in andalusite relative to sillimanite towards the northwest (shallower erosion level) suggests that metamorphic conditions were also at or near that of the reaction

(5) ondolusite = sillimanite.

At P — T conditions for this inversion, the limited free energy of the reaction does not tend to favor one polymorph over the other (Pitcher and Flinn, 1965; Holdaway, 1971) so that one phase may exist metastably. P — T conditions for reactions 4 and 5 are plotted in figure 3.

The univariant boundary for reaction 5 is taken from Holdaway (1971). The position of this boundary, however, varies from one experimental (and probably natural) design to another. The data of Holdaway has been used in this study mainly because he appears to have better control over the metastable inversion of fibrolite to andalusite which seems to have plagued earlier investigators. The $P_f = P_t - T$ conditions indicated by reactions 1, 3, 4, and 5 are approximately 2.0 to 2.5 kb and 600°C. These reactions, however, are all bivariant to some degree depending on grain size, composition of the fluid phase, major and trace element content, etc. but probably represent an effective maximum temperature within the assigned pressure interval. These low pressure-high temperature conditions are similar to those described by Miyashiro (1961) for the Abukuma-type regional metamorphism and by Schwarcz (1969) for the French Valley Formation. The assemblage cordierite-sillimanite-andalusite was used by Miyashiro to indicate metamorphism under a steep geothermal gradient. The geothermal gradient estimated for the Morena Reservoir roof pendant ranges from about 65°C/km to 85°C/km.

CONCLUSIONS

The bedding, as defined by sedimentary structures and by variations in mineralogy and grain size, is everywhere parallel to the metamorphic foliation. Development of the foliation, there-

ire, must have preceded the folding. The porphyroblasts of muscovite that are oriented at high angles to the foliation may hen represent a post-metamorphic introduction of materials elated to the emplacement of the granitic components of the atholith.

Several small satellitic plugs of the adjacent Los Pinos Gabbro re intrusive into the roof pendant (Walawender, 1976). In ddition, an undeformed diabase dike with strongly zoned plagiclase phenocrysts cuts across the axial zone of the antiform. his dike has been shown to be texturally and chemically related o the Los Pinos Gabbro (Berggreen, 1976; Walawender, 1976). he emplacement of the gabbro and the dike, therefore, must ostdate the metamorphism and folding of the metasedimentary ocks. Since the gabbro is the oldest plutonic unit in this area ith a K-Ar age of 143 ± 8 m.y.B.P. (Krummenacher and thers, 1975), the periods of metamorphism and folding must ave also preceded the emplacement of the granitic components f the batholith.

The coarse grain size and high degree of rounding in the clasts f the metaconglomerate combined with the high degree of ounding in the relict cores of the original particles in the schistse units and the presence of minor carbonate in the original ocks suggests deposition in a shallow water environment. Gastil (1975) depicts the Santiago Peak Volcanics as an island arc complex over an eastward-dipping Jurassic-Cretaceous subduction zone. The timing of the development of the Jurassic volcanic rc relative to the metamorphism of the Triassic(?) sedimentary ile, however, has not yet been established. Additional studies on he metamorphic units west of the present study area are necesary before the stratigraphic relationships of the dominantly edimentary and the dominantly volcanic prebatholithic sections an be determined.

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THE PRESTON PEAK OPHIOLITE, KLAMATH MOUNTAINS, CALIFORNIA, AN IMMATURE ISLAND ARC: PETROCHEMICAL EVIDENCE [1]

By ARTHUR W. SNOKE [2], H. R. BOWMAN [3], and A. J. HEBERT [3]

ABSTRACT

Major and trace element contents (including rare earth elements) have been determined on a suite of rocks from the Preston Peak ophiolite, Klamath Mountains, California. These geochemical data combined with field relations and petrography suggest that the Preston Peak ophiolite is a remnant of a Permian-Triassic immature island arc. The lowest unit in the ophiolite is a depleted peridotite tectonite which was subsequently intruded and overlain by mafic rocks as the ophiolite developed. Diabase and diabase breccia, the most common constituents of the mafic complex, have undergone greenschist facies metamorphism and related chemical migration. Nevertheless, low TiO_2 contents as well as low Cr values suggest similarity with island arc tholeiite rather than abyssal tholeiite. Xenoliths of clinopyroxenite and gabbro, locally common in the diabasic rocks, appear to be cumulates developed in small magma chambers that were later disrupted by intruding younger basaltic magmas. Evidence of extensive differentiation is lacking and apparently much of the fine-grained mafic rocks of the ophiolite represent primary magmas. The most leucocratic and siliceous rock in the suite is a Na-rich quartz diorite which is enriched in light rare earth elements and probably represents an independent magma type rather than a differentiate.

INTRODUCTION

Extensive terranes of ophiolitic rocks are widely developed throughout the Klamath Mountains of northwestern California and southwestern Oregon (Coleman and Irwin, 1974; figure 1). The ages of these complexes are uncertain; but minimum ages, based on overlying fossiliferous strata, decrease oceanward (east to west) across the province (Irwin, 1973). The ophiolites are concentrated along the boundaries between major lithic belts (figure 1) and are bounded along their basal contact by a fault or fault system of regional extent (Irwin, 1964; Davis, 1968). The emplacement of the ophiolites is not fully understood although an accretionary phenomenon related to an evolving series of east-dipping subduction zones is often suggested (Hamilton, 1969; Burchfiel and Davis, 1972, 1975; Coleman and Irwin, 1974). In the Preston Peak area, Del Norte and Siskiyou Counties, California (figure 1) a particularly well-exposed ophiolite forms the base of the western Paleozoic and Triassic belt of Irwin (1960). Field relations (Snoke, 1972, in press) indicate that the complex has a pseudo-stratigraphic aspect ranging upward from a basal sheet of peridotite tectonite through a mafic complex chiefly of metadiabase and metadiabase breccia and overlain by metabasaltic and metasedimentary rocks (figure 2).

An important conclusion drawn from field and petrographic studies of the Preston Peak ophiolite (Snoke, 1972, in press) is that the basal tectonitic ultramafic rocks and the overlying constructional pile have undergone different crystallization histories. This is indicated by the sharp metamorphic contrast between the tectonitic ultramafic rocks and the non-schistose lower greenshist facies metadiabase and metagabbro. The local preservation of intrusive contacts between younger mafic rocks and tectonitic peridotite substantiates this hiatus in the development of the ophiolite. Furthermore, widespread mafic dikes with chilled margins indicate that the ultramafic rocks have acted as country rock rather than source rock for the basaltic magmas.

The purpose of the present study was to make a petrochemical reconnaissance of the Preston Peak ophiolite to define its chemical character and variation. These new chemical data along with available field evidence suggest that the ophiolite may represent a remnant of an immature island arc. However, these data are not unequivocal, and the uncertainties of this interpretation are enumerated.

PETROGRAPHY

The analyzed rocks from the Preston Peak ophiolite can be divided into four distinct groups:

1. Serpentinized tectanitic peridatite (alpine-type)—twa analyses (majar and trace elements).

2. Diabase and diabase breccia—six analyses (majar elements), nine analyses (trace elements).

3. Caarse-grained xenaliths in diabase—twa analyses (majar and trace elements).

4. Nanparphyritic biatite quartz diarite—one analysis (majar and trace elements).

Tectonitic Peridotite

The analyzed alpine-type peridotites are massive and unsheared but extensively serpentinized. However, relict primary minerals are common and indicate that the original mineral assemblage was Mg-rich olivine, enstatite, diopsidic augite, and chromian spinel. In addition to serpentine-group minerals, other common alteration minerals are talc, chlorite, tremolite, and magnetite. Trace amounts of carbonate and brucite are also present.

A relict mesoscopic fabric is preserved in these rocks and is defined by a planar arrangement of ellipsoidal aggregates of olivine and pyroxene. In sample PP-225 distinct pyroxene-rich segregation bands (about 5 mm thick) are well developed. The relict primary mineral assemblage and the penetrative fabric indicate that these rocks have undergone high-temperature recrystallization and flowage. Similar ultramafic rocks form the basal part of other Klamath Mountains ophiolites (Lipman, 1964; Himmelberg and Loney, 1973; Goullaud, 1975).

Diabase and Diabase Breccia

Diabase and diabase breccia constitute more than 95 per cent of the mafic complex. Other common components are gabbroic and clinopyroxenite xenoliths (figure 3) and dikes and small masses of aphanitic greenstone. All these rocks are non-schistose but have undergone low-grade greenschist facies metamorphism. The typical mineral assemblage is altered plagioclase, actinolitic amphibole, epidote, chlorite, and leucoxene (e.g., PP-376, figure 3B). Other rocks have been contact metamorphosed by younger and unrelated Late Jurassic plutons.These rocks consist chiefly of oligoclase-andesine plagioclase and green hornblende; they typically lack or contain only minor quantities of chlorite and epidote. Although several diabase specimens have undergone incipient contact metamorphism (PP-317, PP-320, S-13B), all the analyzed diabases retain their original igneous texture and

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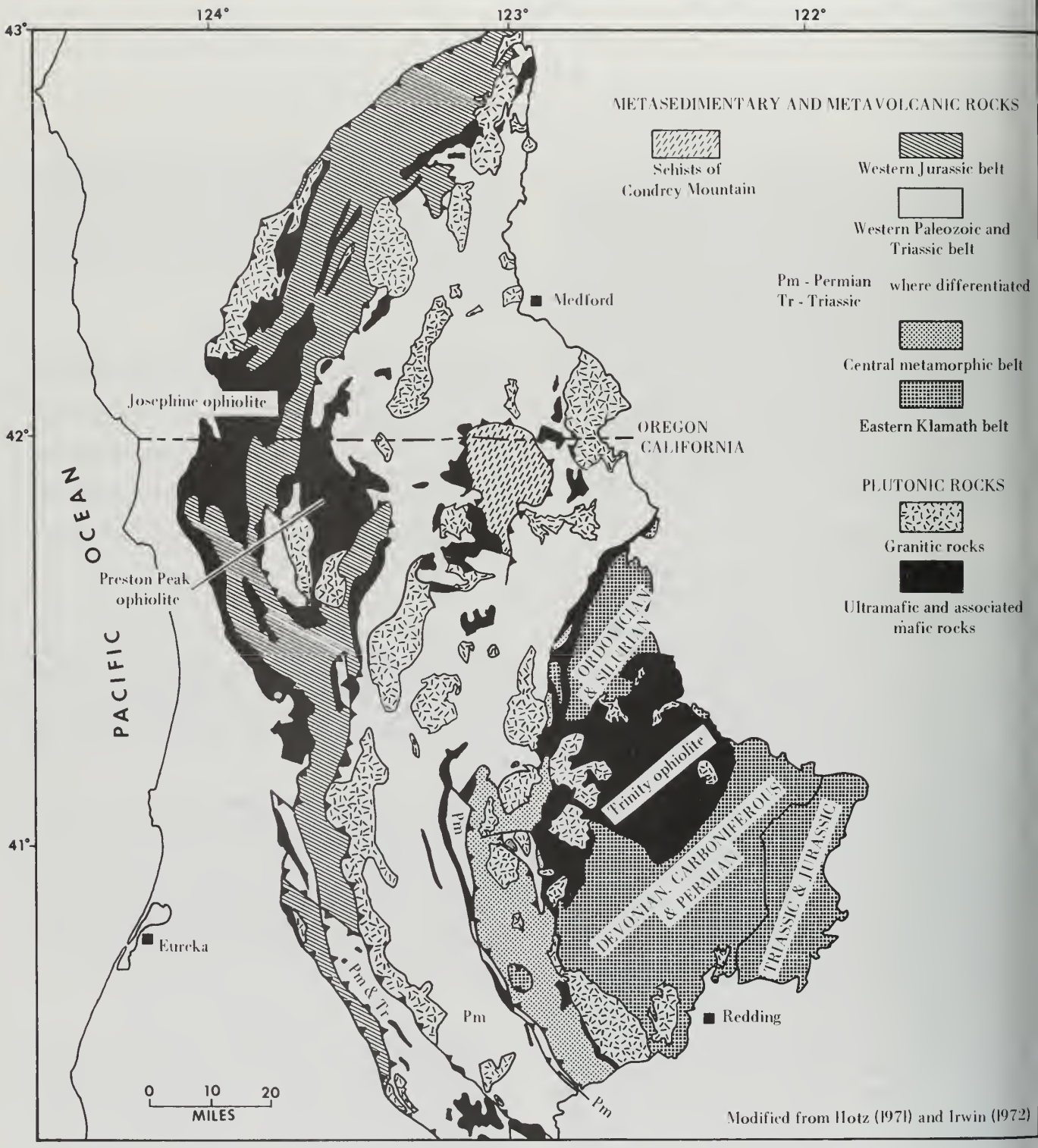
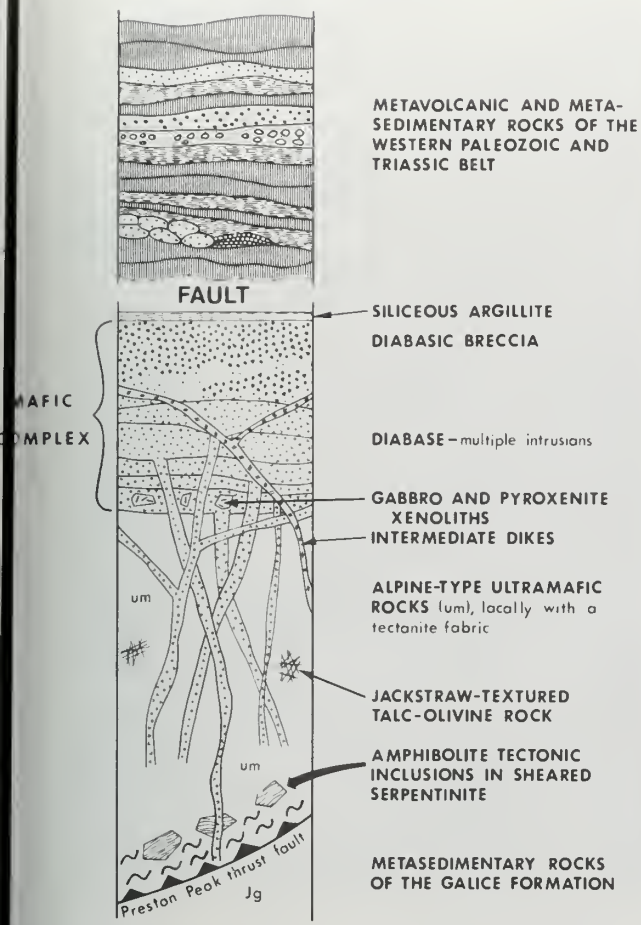


Figure 1. Generalized geologic map of the Klamath Mountains.



igure 2. Schematic stratigraphic column of the Preston Peak ophiolite.

how little indication of extensive grain growth and recrystallization.

The diabase breccia (PP-381) consists of poorly sorted, angular fragments of mafic igneous rock. Diabase and aphanitic greenstone clasts are predominant, but microgabbro as well as porphyritic and vesicular basalt fragments are common. These clasts show no evidence of aqueous transport. The rock unit is interpreted to be pyroclastic breccia erupted from a nearby explosive vent. The coexisting mineral assemblage in the breccia indicates greenschist facies metamorphism.

Coarse-Grained Xenoliths

Gabbroic and less commonly clinopyroxenite xenoliths are found in metadiabasic rocks of the mafic complex. The xenoliths are apparently scattered throughout the mafic complex but appear to be somewhat preferentially concentrated near basal ultramafic-mafic rock contacts.

The xenoliths, like their host rock, have been extensively overprinted by greenschist facies metamorphism. The gabbro originally consisted of calcic plagioclase, augitic pyroxene, and opaque oxides (figure 3C). Plagioclase grains are now heavily saussuritized but retain their original igneous crystal shape. Pyroxene grains are pseudomorphed by fibrous green actinolitic amphibole and irregular chlorite patches. The primary minerals of the pyroxenites are likewise extensively altered. Original olivine grains are completely pseudomorphosed by serpentine minerals, while pyroxenes are typically only partially replaced by amphibole. There is no indication that orthopyroxene was a primary mineral phase in any of the xenoliths.

Biotite Quartz Diorite

Intermediate dikes intrude metadiabase and serpentinized peridotite in several localities throughout the Preston Peak area. If all of these dikes are part of the ophiolite is uncertain, for Late Jurassic magmatic activity is widespread throughout the area. The analyzed biotite quartz diorite (PP-322) has undergone a low-grade metamorphic alteration characteristic of the ophiolitic rocks and is texturally distinct from any known Jurassic igneous rocks. This rock is considered to be an example of the leucocratic suite apparently common in ophiolitic assemblages (Coleman and Peterman, 1975).

The rock consists essentially of strongly zoned subtabular plagioclase (andesine to oligoclase), interstitial quartz, ragged biotite, and opaque minerals (figure 3D). Minor accessories include sphene and apatite. No potash feldspar was seen. Saussuritic alteration of plagioclase cores is common. Similarly biotite aggregates are often partially altered to chlorite and muscovite.

ANALYTICAL METHODS

The major elements were determined by X-ray fluorescence analysis using the method described by Hebert and Street (1973). The minor and trace element abundances were obtained using neutron activation analysis (N.A.A.) and the method has been described by Bowman and others (1973). The N.A.A. data are shown in table 2 and the indicated errors are based on counting statistics and are related more to precision than accuracy. A measure of the overall accuracies of these numbers can be obtained by comparing the present data with our measurements of BCR-1 Standard Rock (Mark and others, 1975). In the BCR-1 data we have included our calibration errors which are introduced by the standards used.

CHEMICAL RELATIONS BETWEEN MAJOR ROCK TYPES

Major element variations as well as chemical relations between the lithologic units of the ophiolite are summarized in figures 4–6. A plot of the analyses on an AFM diagram (figure 4) indicates that the rocks clearly fall into distinct groups compatible with field relations and petrography. A study of major element chemical variation in the Coast Range ophiolite of California (Bailey and Blake, 1974) indicate similar divisions between the major lithologies. A FeO* vs. FeO*/MgO plot (figure 5) shows somewhat similar divisions but also suggests an indication of iron enrichment in diabase (e.g., PP-320). This minor iron enrichment is also detectable on the AFM diagram (figure 4). A SiO_2 vs. FeO*/MgO plot (figure 6) spreads out the diabase group based on their SiO_2 variation (47.9 to 52.4) and suggests slight SiO_2 enrichment, perhaps related to limited magmatic differentiation. As will be developed later in this paper, we argue that evidence of extensive fractional crystallization is not apparent in the ophiolite. Furthermore, PP-322 the most siliceous rock of the analyzed suite does not appear to be a late-stage differentiate but rather a sample of an independent magma type.

CHEMICAL CHARACTERISTICS OF MAJOR ROCK TYPES

Tectonitic Peridotite

As typical of many Cordilleran alpine-type ultramafic rocks (Thayer and Himmelberg, 1968; Himmelberg and Loney, 1973; Bailey and Blake, 1974), the analyzed tectonitic peridotite is characterized by low content of Al_2O_3 and CaO. The low CaO

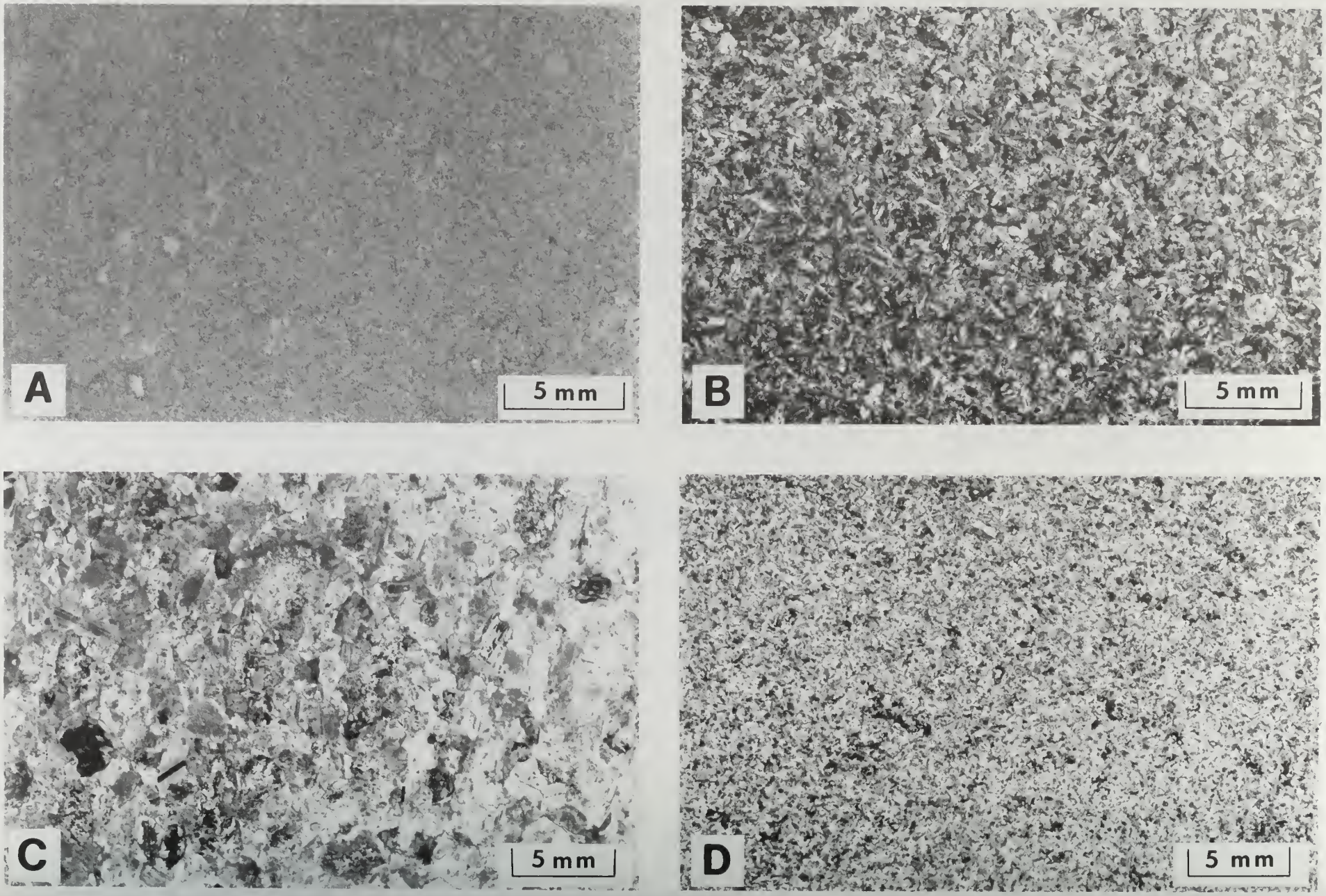


Figure 3. Photomicrographs of analyzed rocks from the Preston Peak ophiolite. (A) Porphyritic metadiabase (PP-384) with chlorite pseudomorphs after pyroxene. (B) Relict pyroxene-bearing metadiabase (PP-376). (C) Metagabbro (PP-374) with relict hypidiomorphic-granular texture. (D) Biotite quartz diorite (PP-322).

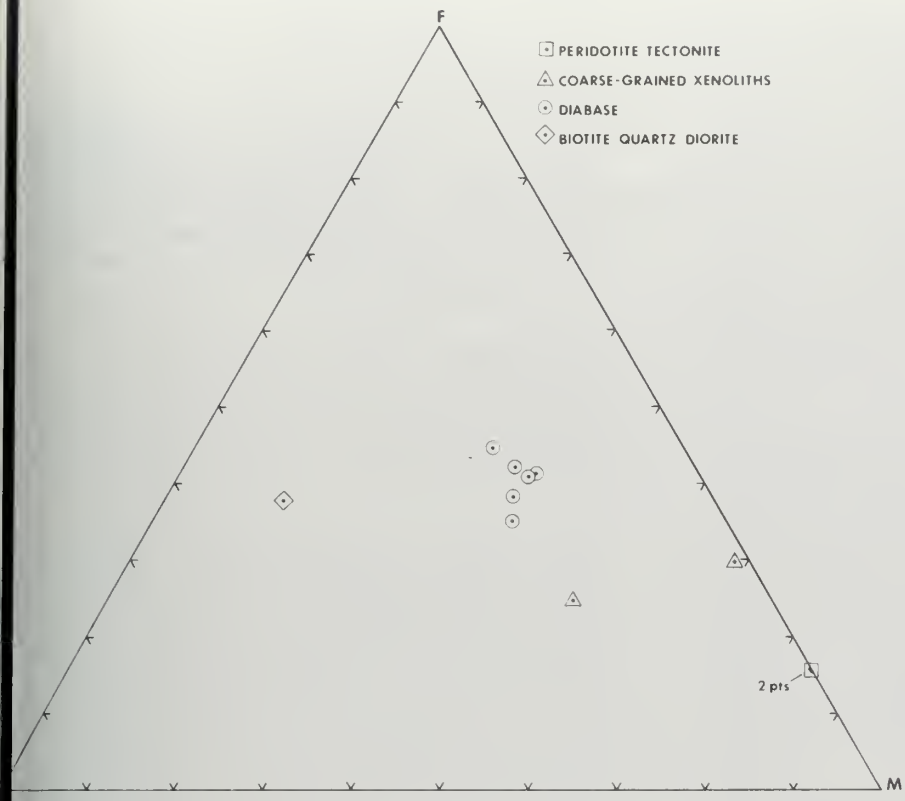


Figure 4. AFM diagram showing peridatite tectanite, coarse-grained xenoliths, diabase, and biotite quartz diarite.

well as the SiO_2 values also reflect the extensive serpentiniza- n that the rock has undergone (Coleman, 1967). When nor- lized on a water-free basis SiO_2 contents are 45.0 and 45.7 for ecimens PP-225 and PP-401A, respectively. Normalized total n to MgO ratios are .188 and .190, respectively. Both samples e very depleted in alkalis as well as almost all trace elements cept Cr and Ni. Minor amounts of Co, V, Zn, and Sc are also esent. These characteristics are in accord with the concept that e peridotite tectonite at the base of ophiolite is a residue re- aining after the removal of basaltic liquids during partial melt- g of primitive upper mantle rock (Coleman, 1971).

Diabase and Diabase Breccia

The restricted range in silica content (~48-52.5) and the high a_2O/K_2O ratios are the most distinctive major element charac- ristics of this rock group. The sodic-rich character of these cks coupled with a wide CaO range (5.8–10.3) are earmarks mafic rocks which have undergone extensive spilitic altera- on. This spilitic character, therefore, makes comparison with altered volcanic suites difficult if not impossible for almost all e major elements. An exception is TiO_2 which is thought to be affected by low temperature alteration (Pearce and Cann, 71, 1973). As indicated in table 1 and displayed on figure 7, O_2 abundances fall within a restricted range (0.70–1.20). Sig- ficantly, this range is clearly within the TiO_2 distribution in mature island arc tholeiite (Jakes and Gill, 1970; Ewart and yan, 1972; Miyashiro, 1974) but is lower than the average ean-floor (abyssal) tholeiite (Cann, 1971).

Another element considered to be useful in distinguishing trotectonic igneous suites is Cr (Bloxam and Lewis, 1972). he average Cr value for the diabase-diabase breccia unit is 135 om (range—34 to 209 ppm). This value is less than typical yssal tholeiite (e.g., Engel and others, 1965, Smewing and hers,1975), and is closer to values given for average island arc tholeiite and calc-alkali basalt (Jakes and White, 1972). When these data are combined with the coexisting TiO_2 abundances, the plotted points form a trend from island arc tholeiite toward abyssal tholeiite (figure 7).

Field relations and petrography of the diabase suite indicate no evidence of major differentiation in place. Furthermore, the flat rare earth element (REE) distribution patterns characteristic of the diabase group (figure 8) also suggest that high-level fractionation was minimal. Such REE patterns are often considered characteristic of primitive magmas developed from partial melting (20 to 30%) of mantle material (Gast, 1968; Kay and others, 1970). However, the coarse-grained xenoliths and indications of slight Fe and/or Si enrichment suggest that differentiation was locally operative during part of the development of the mafic complex. We reconcile these two opposing views by arguing that the bulk of the mafic complex was built of primitive and unfractionated magma; however, local ponding of magma is responsible for the scattered evidence of differentiation. Apparently these small magma chambers were disrupted by younger intrusive basalt or occasionally tapped of their fractionated liquids. Judging from the paucity of coarse-grained xenoliths in the mafic complex, most of the magmas intruded high into the crust and cooled quickly as flows, dikes, or sills.

Many studies (e.g., Frey and others, 1968; Philpotts and others, 1969; Kay and others, 1970) indicate that REE abundances are unaffected by post-magmatic alterations and, therefore, are useful in characterizing magma types. However, substantial data (e.g., Jakes and Gill, 1970; Ewart and Bryan, 1972) also indicate that the flat REE patterns are characteristic of basalt from contrasting tectonic environments (i.e., abyssal tholeiite *vs.* island arc basalt). Therefore, although our REE data unquestionably suggest a tholeiitic protolith for the metadiabase, we cannot unequivocally indicate the site of origin. Kay and Senechal (1976) likewise pointed out this problem in their analysis of REE data on basalt from the Troodos ophiolite complex.

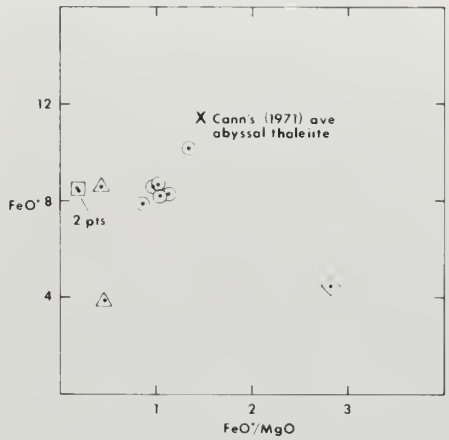
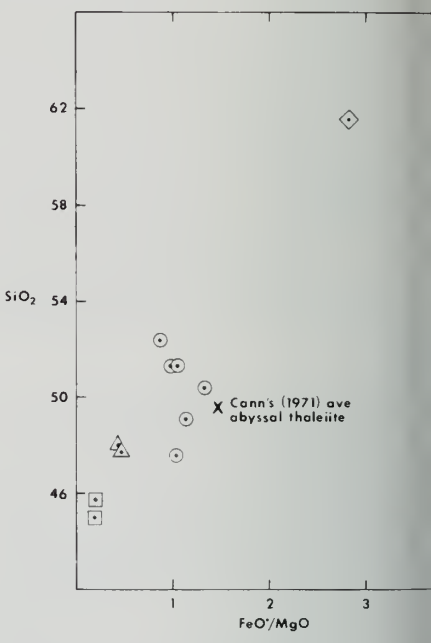


Figure 5. FeO **versus* FeO */MgO plot showing peridatite tectonite, caarse-grained xenoliths, diabase, and biatite quartz diarite. Rock types are designated with the same symbals as in figure 4. Conn's (1971) average obyssol tholeiite is shawn for camparisan.

Figure 6. SiO *versus* FeO */MgO plot showing peridatite tectonite, caarse-grained xenaliths, diabase, ond biatite quartz diarite. Rock types are designoted with the same symbols os in figure 4. Cann's (1971) average obyssal thaleiite is shawn far camparisan.



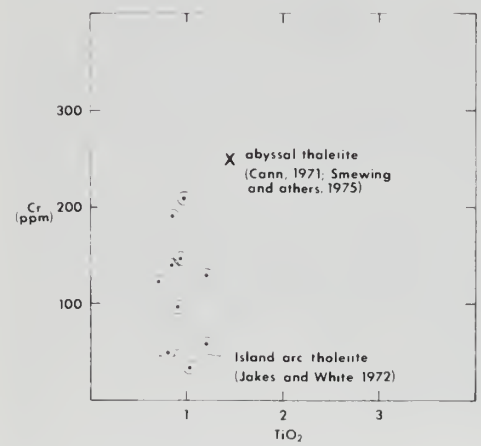


Figure 7. Cr *versus* TiO_2 plot far Preston Peok diobosic rocks, circles. Averoge volues far obyssol tholeiite ond island orc tholeiite ore platted for comparisan.

Coarse-Grained Xenoliths

Although only two specimens of xenoliths were analyzed, the noliths appear to constitute a chemically distinct group within ophiolite suite. In regard to major element content this is ecially apparent in the AFM diagram (figure 4) on which the nolith specimens plot toward the MgO corner. Likewise in ures 5 and 6, the xenoliths have distinctly lower FeO*/MgO ios than the other rocks with the exception of the peridotite tonite. These data coupled with petrography and field occurnce suggest that these rocks are cumulates brought up by unger basaltic magmas.

The clinopyroxenite specimen (PP-382) in comparison to the ine-type peridotite tectonites has distinctly higher Al_2O_3, CaO d TiO_2 contents, reflecting the abundance of clinopyroxene. e gabbro xenolith (PP-374) in comparison to the diabasic ks is characterized by a higher Al_2O_3 and CaO content and ks the high Na_2O/K_2O ratio.

The low REE contents of these rocks is consistent with a mulate origin; however, the analyzed specimens are not as pleted in REE as many gabbros from the Troodos complex ay and Senechal, 1976) or from the Pindos suite (Montigny d others, 1973). Perhaps as suggested by Kay and Senechal 976) for a similar Troodos gabbro, the Preston Peak xenoliths adcumulates which once contained a significant quantity of pped liquid. According to Kay and Senechal (1976), the pped liquid would contribute most of the REE. Therefore, cause REE patterns from xenoliths and diabase are crudely rallel, we argue that both groups were derived from magmas th similar REE concentrations.

Biotite Quartz Diorite

The biotite quartz diorite specimen (PP-322) is the most silious of the analyzed rocks from the Preston Peak ophiolite, and has the highest FeO*/MgO ratio. It is characterized by a high a_2O/K_2O ratio but also contains substantial CaO. These characteristics are typical of rocks of the keratophyre-trondhjemite ite, often an integral part of many ophiolite complexes (Coleman and Peterman, 1975).

The REE pattern (figure 8) for the quartz diorite (PP-322) ows enrichment in light REE (~50 times the Leedey chonite) with a depletion of heavy REE. No Eu anomaly is apparnt. This pattern is strikingly different from Troodos agiogranite (Coleman and Peterman, 1975; Kay and Senechal, 976) and appears similar to fractionated REE patterns in the recambrian Saganaga tonalite and the Mesozoic Craggy Peak ndhjemite as reported by Arth and Hanson (1972). Accordg to Arth and Hanson (1972), such patterns which show a pletion of heavy REE are consistent with an origin involving rtial melting of amphibolite or eclogite at depths greater than km. Barker and others (1976) have also documented Precamian trondhjemitic rocks ($Al_2O_3 > 15$ wt%) with similar REE tterns. They favor a partial melting mechanism at greater than km but point out that rocks with similar REE patterns could produced by fractional crystallization at crustal levels. Reardless of the origin, our data indicate that leucocratic rocks sociated with ophiolite may be more variable than previously mphasized (e.g., Coleman and Peterman, 1975).

EVALUATION OF THE PETROGENESIS OF THE PRESTON PEAK OPHIOLITE

Despite substantial new data on ophiolite in recent years, the te of origin of this enigmatic rock sequence remains a controversial problem in petrology. At present, three models are commonly considered for the origin of ophiolite complexes:

1. They formed at on oceanic ridge in a majar oceon bosin (Coleman, 1971; Dewey and Bird, 1971; Moores and Vine, 1971).

2. They farmed in o marginal bosin behind an active island arc (Dewey and Bird, 1971; Blake and Jones, 1974; Dewey, 1974).

3. They represent the immoture stoge (i.e., the bosol part) of an islond orc complex (Ewart and Bryan, 1972; Miyoshira, 1973, 1975).

We concur with Kay and Senechal (1976) that whole rock chemical analyses including major and trace element data cannot unequivocally discriminate between the above three models. However, we hasten to add that our chemical data do indicate certain constraints for each lithologic group of the Preston Peak ophiolite:

1. The ultromafic tectonite is chemically and texturolly distinct from all other rocks in the ophiolite. Its depleted chorocter caupled with its metamorphic texture suggests thot it is residual montle rack.

2. The diabasic rocks show only minor froctionotian and prabably represent primory mogmos that hove crystallized in a subvolconic environment. The diobase breccio unit is chemically similar ta the massive, nonfrogmentol diobase and probably is o pyroclastic equivolent.

3. Chemicol doto coupled with field occurrence and petrography suggest that the coorse-groined xenoliths are cumulotes rafted up by younger basoltic mogmos. REE dato indicote thot they may be derived fram mogmos similor in composition to their enclosing diobasic hosts.

4. The fractionoted REE pattern of the biotite quartz diarite is analogaus to certain No-rich calc-alkoline rocks. Therefore, this rack type probably represents o distinct mogmo type chemically unreloted to the mafic rocks of the ophiolite.

Detailed evaluation of the genesis of the Preston Peak ophiolite as deduced from field data is presented elsewhere (Snoke, in press). We will not reiterate these statements but emphasize that many observations seem incompatible with known data and postulated models for modern spreading centers either in marginal or major ocean basins. Some of the key points are:

1. Metadiabase dikes in the basal peridotite appear ta be feeders that in part nourished the overlying subvalcanic mafic complex. These dikes form sharp contacts with chilled selvages adjacent to tectonitic ultramafic rocks. There is no evidence, such as magmatic segregations (Dickey, 1970), that the mafic magmas were derived fram the surraunding depleted peridatite.

2. The tectonite fabric (i.e., metamorphic recrystallization) of the basal peridotite predates mast af the mafic magmatism of the Preston Peak ophiolite. An exception is the amphibolite masses which occur in serpentinite melange along the basal fault contact (figure 2). These racks may be remnants of oceanic crust upon which the mafic camplex was built.

3. Mafic-ultramafic cumulate rocks, charocteristic of many ophiolites and cammonly postulated to be abundant at spreading centers (e.g., Moores and Jackson, 1974; Jackson and others, 1975), are scarce in the Preston Peak aphiolite. Perhaps this feature is indicative of a fast spreading ridge (Moores and Vine, 1971), but alternatively it may represent a fundamental difference in site af origin.

4. Recent investigations of Mid-Atlantic Ridge volcanism (Moore and athers, 1974) indicate that a preponderance af pillow lava erupted quietly from fissure vents. Pillaw lava is nat abundant in the Preston Peak aphiolite; in fact pyraclastic mafic breccia is common and probably is evidence of local explosive volcanic centers.

These observations as well as certain enumerated aspects of the petrochemistry suggest to us that the young island arc model is a possible alternative. We can only speculate on what factors controlled the initiation of the proposed Preston Peak arc but perhaps late Paleozoic geanticlinal buckling as postulated by Burchfiel and Davis (1975) for the Trinity ophiolite is a viable explanation. Nevertheless, upper mantle peridotite tectonite and metamorphosed oceanic crust (amphibolite) served as the foundation upon which the arc was built. Initial magmas were basaltic (island arc tholeiite) and as the volcanic pile grew, younger magmas often didn't reach the surface but formed an extensive dike-sill complex. The fluidity and probable low pH_2O of the magma precluded extensive pyroclastic deposits during this

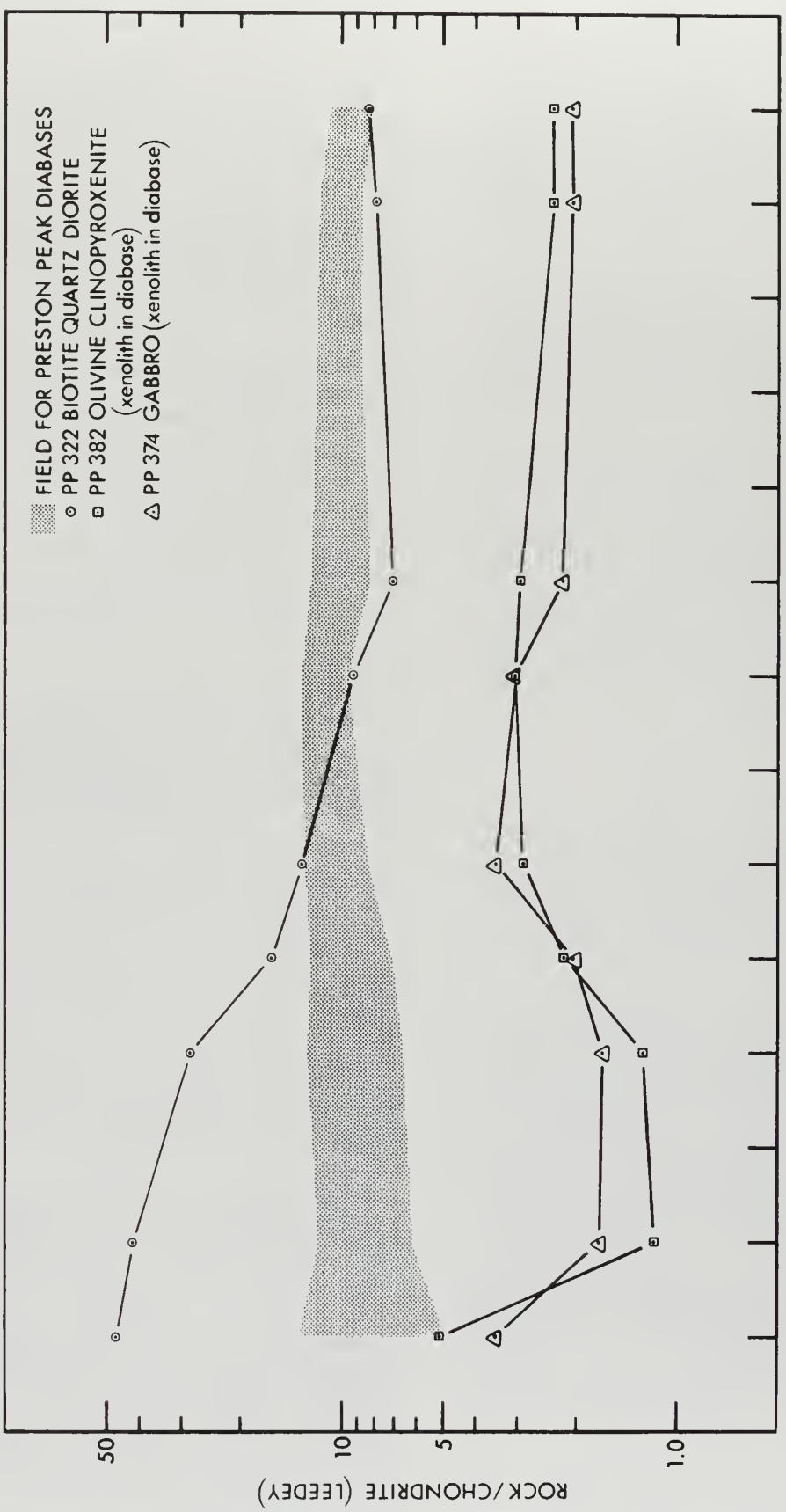
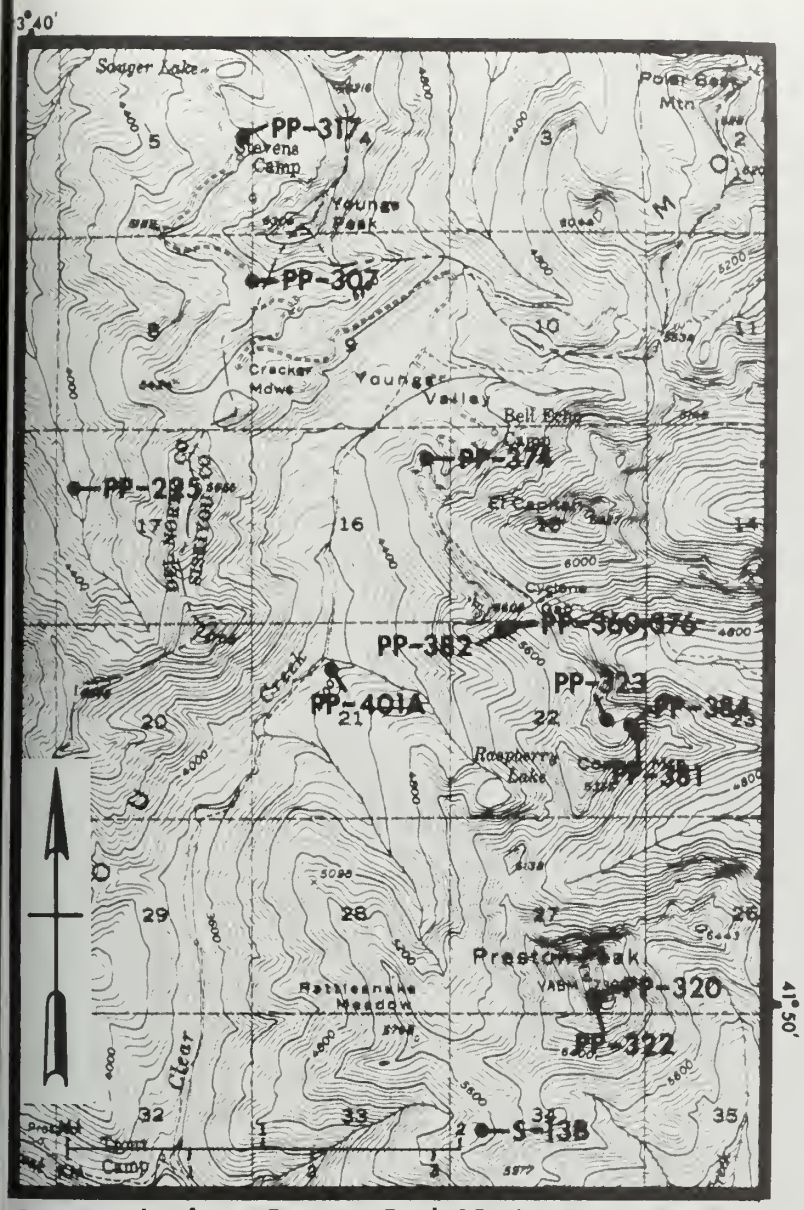


Figure 8. Leedey normalized REE patterns for coarse-grained xenoliths, diabases, and biotite quartz diorite from the Preston Peak ophialite.



Topography from Preston Peak 15min. quadrangle (1956)--U.S.G.S.

Figure 9. Map shawing location of chemically analyzed samples.

age, but local vent breccia occasionally developed. Small mag- chambers formed but commonly were disrupted by younger trusive magmas. As the arc matured, intermediate magmas presented by Na-rich quartz diorite became part of the igneous mplex. Overlying coarse-grained, immature sedimentary cks suggest that local relief became progressively more promi- nt as the arc matured. Intermediate hypabyssal and plutonic agments are phenoclasts in these detrital rocks suggesting that lc-alkaline rocks in the Preston Peak ophiolite were more xtensive than the present exposures indicate (Snoke, in press).

The proposed island arc model appears consistent with the vailable data on the Preston Peak ophiolite, and we know of no egional data to preclude our interpretation. In fact, we note that pparent calc-alkaline igneous rocks appear to form the upper arts of several other ophiolites in the Klamath Mountains (e.g., he Josephine and Trinity ophiolites). The Rogue Formation, escribed by Wells and Walker (1953) as predominantly meta-andesitic flows and breccias, structurally and locally depositionally overlies in the Josephine ophiolite in southwestern Oregon. Analogous volcanic rocks forming the base of the Galice Formation appear to overlie the southern part of the Josephine ophiolite in California (Cater and Wells, 1953). In the southeastern Klamath Mountains, the southern part of the Trinity ophiolite is overlain by the Copley Greenstone, metabasalt to meta-andesite flows and breccias, and the Balaklala Rhyolite, predominantly keratophyre to quartz keratophyre flows and pyroclastic rocks (Kinkel and others, 1956). Therefore, we suggest that these ophiolites should be investigated as possible remnants of the underpinnings of an island arc complex.

ACKNOWLEDGMENTS

We thank Akiho Miyashiro and Eldridge M. Moores for their critical reviews of the manuscript.

Table 1. Partial chemical analyses of Preston Peak ophiolitic rocks *(in percents)*.

	Tectonitic Peridotite				Olivine Clino-pyroxenite	Gabbro	Diabase						Quartz Diorite
	PP-225		PP-401A		PP-382	PP-374	PP-317	PP-320	PP-323	PP-360	PP-376	PP-384	PP-322
SiO	39.7	(45.0)#	38.8	(45.7)	48.0	47.8	51.3	50.4	51.3	47.9	49.1	52.4	61.6
TiO_2*	<.02	(.02)	<.02	(.02)	.12	.20	.97	1.20	.92	.90	.83	.85	.35
Al_2O_3	.19	(.22)	.39	(.46)	4.8	21.8	15.2	15.4	15.2	16.8	15.6	14.9	17.3
FeO*	7.5	(8.5)	7.1	(8.4)	8.6	3.9	8.6	10.2	8.2	8.7	8.3	7.9	4.5
MgO	40.0	(45.3)	37.6	(44.3)	19.8	8.3	8.8	7.7	7.9	8.5	7.4	9.2	1.6
CaO	.81	(.92)	.99	(1.2)	13.2	12.5	7.9	7.0	10.3	10.2	10.3	5.8	5.3
Na_2O*	.02	(.02)	.02	(.02)	.53	2.8	5.0	4.4	3.8	3.7	3.9	5.3	4.7
K_2O	n.d.		n.d.		.01	.73	.08	.56	.11	.23	.12	.15	1.1
Total	88.24		84.92		95.06	98.03	97.85	96.86	97.73	96.93	95.55	96.50	96.45

FeO = Total iron.

n.d. = not detected.

= The number in parentheses is a recalculated value so as to sum 100%.

* = Instrumental neutron activation analysis (Analyst - H.R. Bowman) otherwise X-ray fluorescence (Analyst - A.J. Hebert).

Table 2. Trace element compositions of Preston Peak ophiolitic rocks *(in parts per million)*.

	Tectonitic Peridotite		Olivine Clino-pyroxenite	Gabbro	Diabase								Diabase Breccia	Quartz Diorite
	PP-225	PP-401A	PP-382	PP-374	PP-307	PP-317	PP-320	PP-323	PP-360	PP-376	PP-384	S-13B	PP-381	PP-322
Ba	<12	9±7	30±11	53±9	55±11	68±13	70±13	64±12	89±13	73±13	56±12	11±11	80±12	437±24
Co	113±2	102±1	92±1	26.4±.4	38.4±.6	38.6±.6	38.1±.6	38.7±.6	41.3±.6	38.5±.6	37.3±.6	43.0±.7	45.3±.7	2.5±.1
Cr	2602±91	2663±93	4298±150	649±23	34±3	209±8	59±3	147±6	98±4	140±5	191±7	123±5	130±5	2.3±1.3
Hf	<.12	<.12	.12±.10	.23±.07	1.9±.2	1.8±.2	2.1±.2	1.7±.1	1.9±.2	1.8±.2	1.6±.1	1.3±.1	2.0±.2	3.5±.3
Mn	1547±49	859±27	1205±38	573±18	1359±43	1276±41	1419±46	1302±41	1444±47	1244±40	1093±35	1250±40	7717±234	1008±32
Ni	2262±176	2259±177	914±75	165±20	32±18	101±21	55±20	72±18	75±18	74±21	84±19	84±19	124±20	5±9
Sc	9.4±.2	9.2±.2	53.0±.9	32.5±.5	39.0±.6	41.3±.7	40.6±.7	39.6±.6	40.2±.6	40.6±.7	36.7±.6	42.2±.7	38.4±.6	7.4±.1
V	64±27	44±11	168±30	119±26	279±45	291±51	405±69	295±51	288±53	324±54	291±49	322±59	173±44	66±19
Zn	69±12	74±12	54±12	13±6	63±11	94±15	82±13	64±11	98±15	86±14	60±11	64±12	115±17	88±12
La	<.5	<.5	1.9±.7	1.3±.6	3.9±.7	3.4±.5	3.7±.5	2.6±.5	6.4±.5	4.4±.5	2.9±.5	1.8±.5	5.0±.5	17.7±.9
Ce	<1.0	<1.0	1.1±.8	1.7±.5	10.7±.6	9.4±.4	9.4±.6	8.3±.5	14.5±.6	12.0±.6	7.6±.5	6.0±.5	11.5±.6	40.9±1.2
Nd	<1.4	<1.6	.9±1.2	1.2±.8	8.6±1.3	7.2±1.5	8.7±1.5	5.7±1.3	7.3±1.5	7.9±1.5	6.2±1.3	4.7±1.2	9.1±1.4	20.1±2.0
Sm	<.006	<.006	.50±.01	.47±.01	2.60±.06	2.29±.05	2.60±.05	2.13±.04	2.72±.06	2.44±.05	2.01±.04	1.63±.03	2.84±.06	3.69±.08
Eu	<.02	<.02	.25±.01	.30±.01	1.02±.03	.87±.03	1.00±.03	.90±.03	.95±.03	.92±.03	.77±.02	.56±.06	1.12±.03	1.13±.03
Tb	<.08	<.08	.18±.07	.18±.05	.61±.06	.63±.06	.72±.06	.65±.06	.65±.06	.61±.06	.57±.06	.56±.06	.77±.06	.55±.04
Dy	<.3	<.2	1.14±.15	.86±.12	4.09±.23	3.75±.23	4.33±.25	3.58±.21	3.52±.23	3.47±.21	3.27±.20	3.19±.20	4.79±.68	2.73±.19
Yb	<.04	<.04	.58±.03	.51±.02	2.62±.06	2.32±.06	2.70±.06	2.30±.06	2.22±.05	2.19±.05	2.16±.05	2.23±.05	2.90±.07	1.97±.05
Lu	<.02	<.02	.09±.02	.08±.01	.37±.02	.35±.02	.37±.02	.34±.02	.33±.02	.30±.02	.31±.02	.33±.02	.42±.02	.32±.02
EREE	3.4	3.5	6.64	6.60	34.51	30.21	33.52	26.50	38.59	34.23	25.79	21.16	38.44	89.09

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DDLE EOCENE MOLLUSCAN ASSEMBLAGE AND STRATIGRAPHY, WER PIRU CREEK, TRANSVERSE RANGES, CALIFORNIA[1]

ICHARD L. SQUIRES[2]

ABSTRACT

previously undescribed middle Eocene molluscon ossembloge composed species (24 gostropod, 17 bivolve, 1 scophopod, 1 corol, and 1 id) and scottered burrows ond plont frogments, occurs in on unnomed tone unit in the lower Piru Creek orea, 10 km north of Piru, Californio. ixed ossembloge, most of which occurs in concretions neor the base of 00–m thick sondstone unit, is interpreted to be o shollow woter, neor- morine environment offected to some degree by current tronsport. rs ore restricted to neor the upper contoct of the sondstone unit, which sents shallower woter conditions. The sondstone unit is grodotionol with derlying offshore siltstone unit ond on overlying nonmorine orkose unit. gradotionol sequence is indicotive of regressive sedimentotion in the during the middle Eocene.

INTRODUCTION

This paper provides a paleontologic analysis of a previously undescribed molluscan assemblage from an area between Piru and Canton Creeks (fig. 1). The molluscan assemblage is indicative of middle Eocene age and occurs in an unnamed sandstone unit deposited in a nearshore, marine environment. This information, some of which was mentioned by Squires (1976), aids in a better understanding of the faunal ranges of the involved taxa and fills an important gap in the paleogeography and sedimentation history of this area during middle Eocene time.

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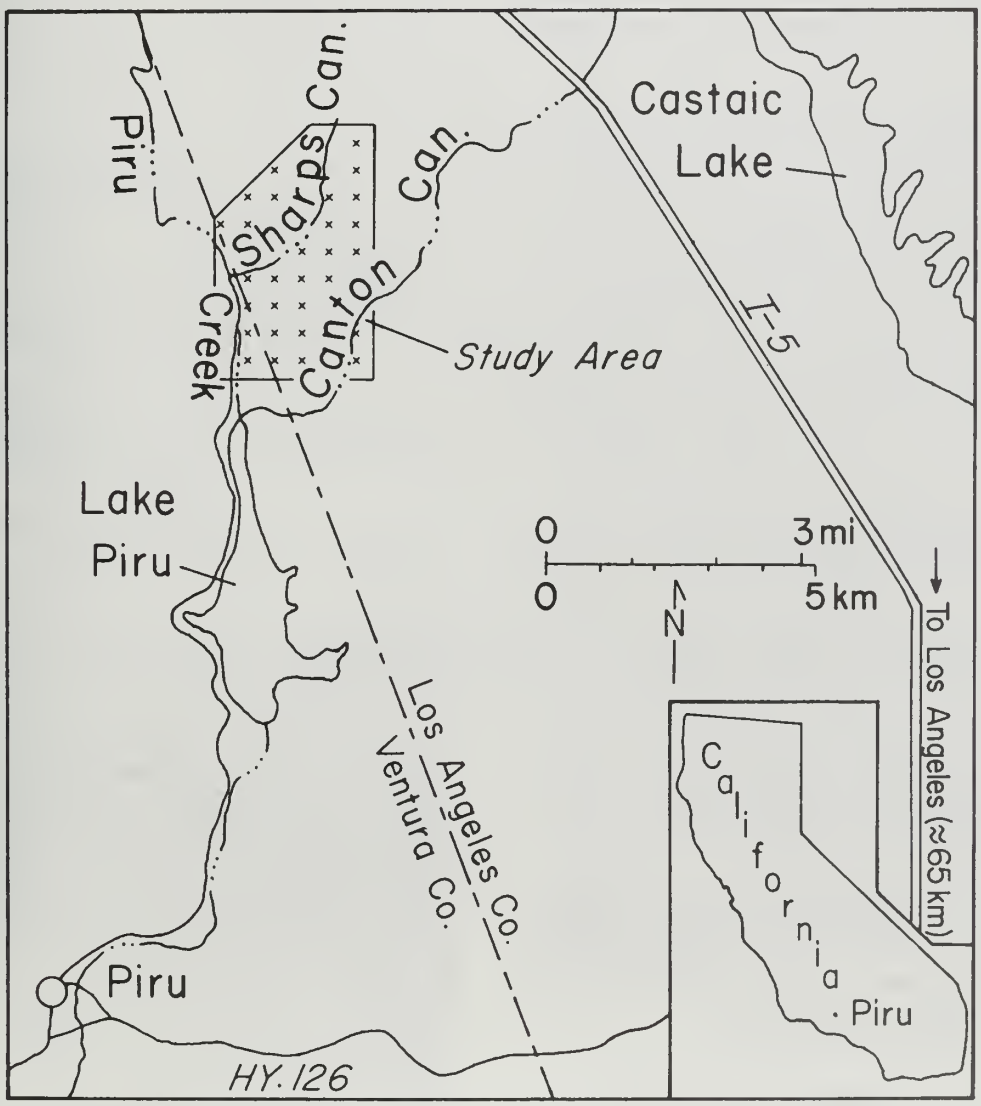


Figure 1. Index map of the lower Piru Creek area.

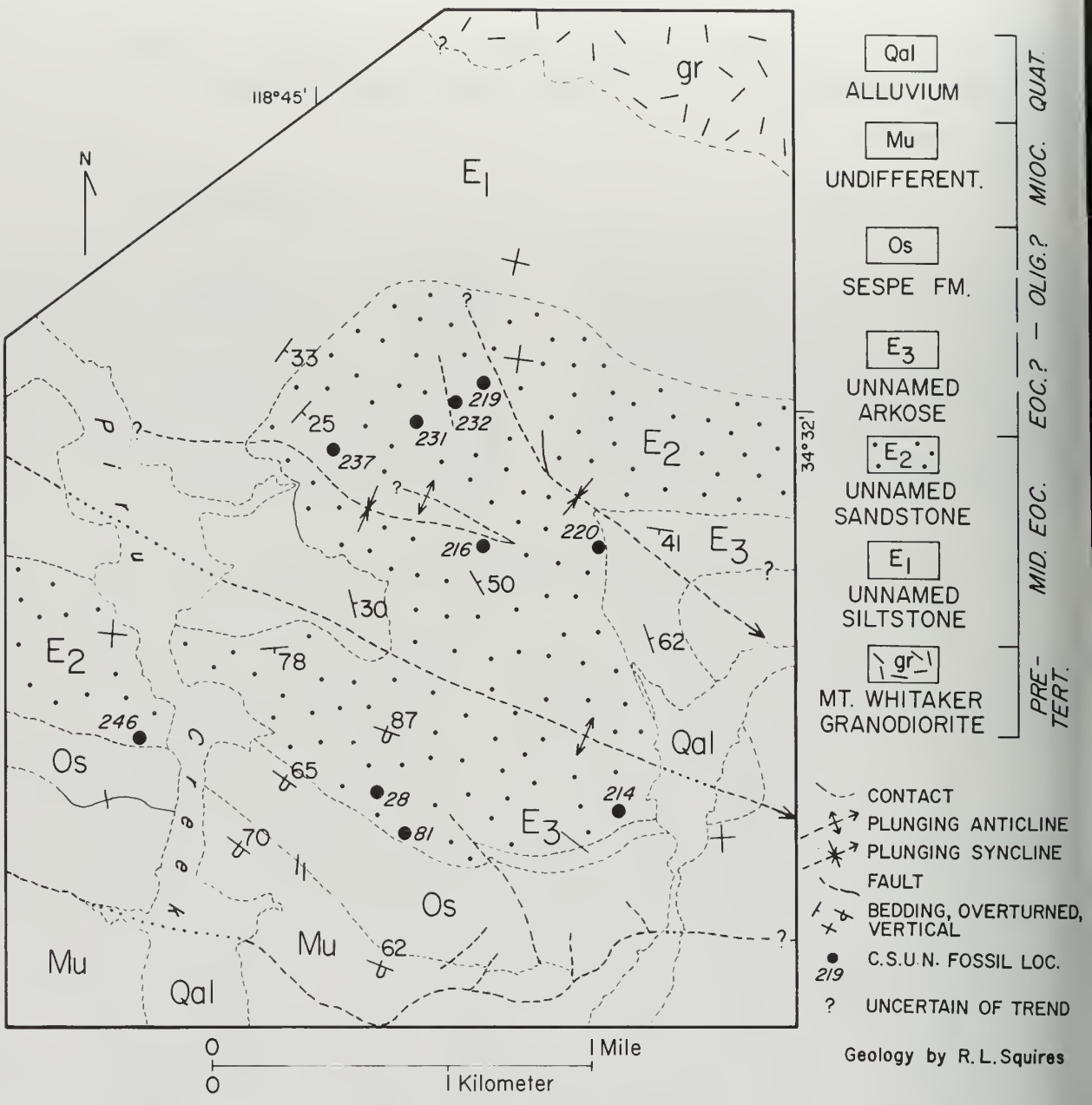
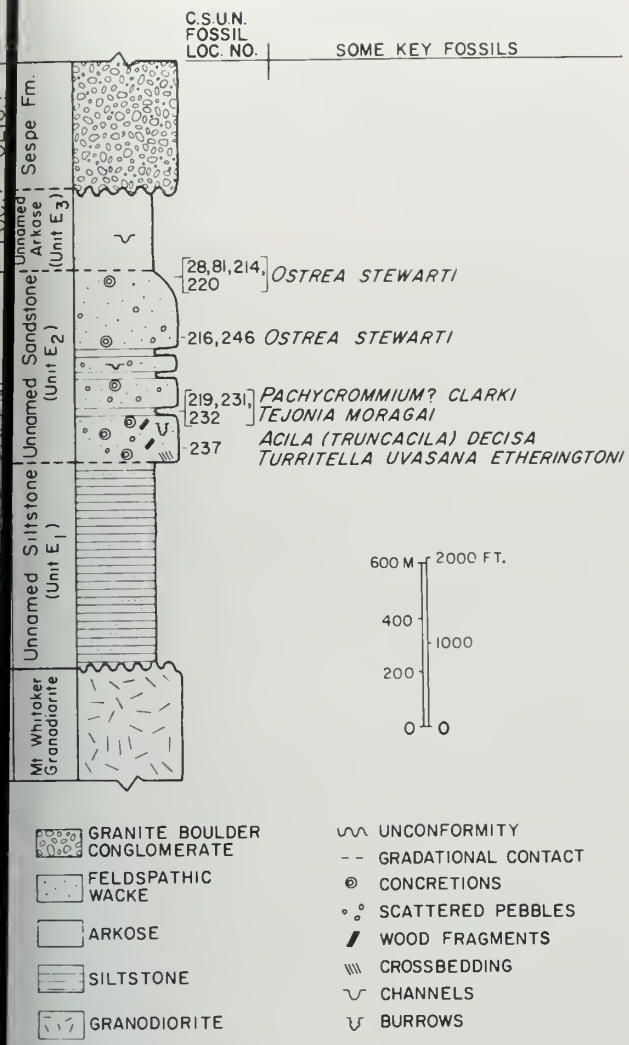


Figure 2. Geologic map of the lower Piru Creek area, Transverse Ranges, California.

GEOLOGY AND GENERAL STRATIGRAPHY

No published detailed geologic maps or reports deal specifically with this area. Previous mapping includes these maps by Kirz (1947), Anderson (1960), Shepherd (1960), and Jestes (1963). A simplified geologic map by Clements (1937) covers the area, and a simplified geologic map by Crowell (1954) includes the extreme eastern part of the area.

Figure 2 is a geologic map by the author. The symbols used for the following mappable units correspond to those on the geologic map. The sequence of rocks is Mt. Whitaker granodiorite, unnamed siltstone (unit E_1), unnamed sandstone (unit E_2), unnamed arkose (unit E_3), Sespe Formation, undifferentiated strata of Miocene age, and Quaternary alluvium. According to Shepherd (1960), unit E_1, 730–m thick, contains middle Eocene megafossils and is essentially unconformable and in most places in fault contact with the granodiorite. Howell (1974) mentioned Eocene foraminifera from unit E_1. Unit E_1 is interpreted in the present report to be an offshore deposit due to its fine grain size. Molluscan fossils from unit E_2 are the focus of this paper and are also indicative of middle Eocene age. Overlying and in gradational contact with unit E_2 is the 300–m thick arkose which has not yielded fossils. Overlying unit E_3 with moderate angular unconformity is the nonmarine Sespe Formation.



ure 3. Stratigraphic column of the Eocene units in the lower Piru ek area.

STRATIGRAPHIC DETAILS OF UNIT E_2

The unnamed sandstone (unit E_2) ranges in thickness from 0 to 700 m and was eroded prior to deposition of the Sespe. s contact is gradational with the underlying siltstone (unit E_1). is mainly whitish yellow, thick-bedded, feldspathic wacke, but ntains scattered pebbles and some interbedded siltstone lenses igure 3). The pebbles are mostly quartzite and granite. In aces, especially near the middle of the unit, the scattered peb- es grade into conglomeratic sandstone beds, some of which ccur within scour channels. The lower part of unit E_2 is charac- ristically closely jointed and the joint surfaces are stained with ematite. Crossbedding, burrows, and carbonaceous zones are rominent locally near the base of unit E_2. Some of the brown ltstone beds in the lower part of unit E_2 are locally fossiliferous nd yield abundant gastropods and bivalves. Most of these fossils ccur in calcareous sandstone concretions, which are 0.3 to 1 m diameter. Prominent purplish-red siltstone beds in the middle f unit E_2 are not fossiliferous. Where the gradational contact ith the overlying arkose occurs, *Ostrea stewarti* specimens are undant just below the contact. Elsewhere, specimens of *Ostrea tewarti* occur in the upper part of unit E_2.

PALEONTOLOGY

The faunal assemblage of unit E_2 consists of 24 gastropod, 17 bivalve, 1 scaphopod, 1 coral, and 1 annelid species. Numerous horizontal and vertical burrows are also preserved in the strata. In addition to carbonaceous layers, one plant fragment 30 cm long was found in a concretion about 50 m above the base of unit E_2.

With the exception of specimens of *Ostrea stewarti* and a few poorly preserved *Tellina* sp., all the fossils were collected in the lower part of unit E_2. C.S.U.N. fossil locality 237 (figure 2), about 50 m above the base of unit E_2, contains the most abundant fossils of all localities (table 1). The fossils occur in a poorly exposed, richly fossiliferous calcareous bed about 1 m thick. Preservation is fair and the fossils can be easily removed from the matrix. Key fossils of this bed include abundant *Turritella uvasana etheringtoni, Acila (Truncacila) decisa, Callista (Microcallista) andersoni, Glycymeris (Glycymeris) rosecanyonensis,* and *Rotularia tejonense.* All specimens of these species display growth series. The gastropod specimens are complete, whereas the bivalves are disarticulated. Several specimens of *Callista (Microcallista) andersoni* have predator boreholes. A coral is represented by a single specimen of *Flabellum* sp. at this locality.

C.S.U.N. fossil localities 219, 231, and 232 (figure 2) are a locally fossiliferous siltstone stratum about 170 m above the base of unit E_2; locality 219 yielded the most abundant and best preserved fossils (table 1). Within this stratum the mollusks occur in calcareous concretions. Most of the concretions are badly weathered except for a few well indurated ones at locality 219, which accounts for the fair to good preservation. Key fossils from this stratum include *Cylichnina tantilla, Pachycrommium? clarki, Tejonia moragai, Turritella uvasana etheringtoni,* and *Nemocardium (Nemocardium) linteum.* All specimens of these species display growth series. Gastropod specimens, all of which are complete, outnumber the bivalve specimens, all of which are disarticulated. Several specimens of *Tejonia moragai* have predator boreholes.

The fossil localities C.S.U.N. 28, 81, 214, and 220 (figure 2) occur in the upper part of unit E_2 and consist almost exclusively of abundant disarticulated specimens of *Ostrea stewarti.* All of the specimens from these localities occur as float material, but the specimens are too abundant, too well preserved, and too predictable in their occurrence to have been derived from another rock unit. Localities 214 and 220 are near the contact between units E_2 and E_3. Localities 28 and 81 are probably at about the same stratigraphic horizon, but due to the unconformable overlap of the Sespe Formation an exact correlation cannot be made. The other fossil localities from the upper part of unit E_2, C.S.U.N. 216 and 246 (figure 2), occur at about 450 m above the base of unit E_2. A single, large, complete specimen of *Ostrea stewarti* was the only fossil found at locality 216. Abundant, large, single valves of *Ostrea stewarti* and a few articulated specimens of *Ostrea stewarti* and *Tellina* sp. were found at locality 246.

PALEOECOLOGY

The gastropod and annelid specimens in the lower part of unit E_2 are unabraded. Many of the specimens are *in situ* forms,

Table 1. Checklist and stratigraphic distribution of megafossils from the unnamed sandstone (unit E_2), lower Piru Creek area.

MEGAFAUNA	Localities			
BIVALVIA	219	231	232	237
Acila (Truncacila) decisa (Conrad)	.....	.....	.....	A
Brachidontes (Brachidontes) cowlitzensis (Weaver & Palmer)	.....	.....	.....	R
Callista (Macrocallista) andersoni (Dickerson)	.....	.....	.....	A
Callista (Microcallista) domenginica Vokes	.....	.....	.....	C
Corbula (Caryocorbula) parilis Gabb	.....	.....	.....	C
Corbula sp.	.....	.....	R	C
Crassatella uvasana semidentata (Cooper)	R	R	.....	R
Gari hornii (Gabb)	.....	.....	F	.....
Glycymeris (Glycymerisa) rosecanyonensis Hanna	.....	.....	.....	A
Glycymeris (Glycymerita) sagittata (Gabb)	.....	.....	.....	C
Glyptoactis domenginica (Vokes)	.....	R	F	R
Nemocardium (Nemocardium) linteum (Conrad)	F	.....	F	.....
Ostrea stewarti Hanna	Only at 28,81,214,216,220,246: A			
Pelecyora gabbi (Arnold)	R	R	.....	F
Pitar sp.	.....	.....	.....	F
Tellina cf. *T. soledadensis* Hanna	.....	.....	R	.....
Tellina sp.	Only at 246: F			
GASTROPODA				
Bonellitia (Admetula) paucivaricata (Gabb)	.....	.....	R	.....
Cerithium cliffensis Hanna	R	.....	.....	.....
Coalingodea tuberculiformis (Hanna)	R	.....	.....	.....
Conus hornii Gabb	R	.....	.....	C
Conus remondii Gabb	R	R	.....	.....
Conus cf. *C. remondii* Gabb	R	.....	.....	.....
Crepidula sp.	R	.....	.....	.....
Cylichnina tantilla (Anderson & Hanna)	C	R	.....	C
Ectinochilus (Macilentos) macilentus (White)	R	.....	R	R
Euspira nuciformis (Gabb)	R	.....	.....	C
Ficopsis remondii crescentensis Weaver & Palmer	R	.....	R	R
Globularia (Eocernina) hannibali (Dickerson)	.....	.....	R	C
Natica (Natica) uvasana Gabb	.....	.....	.....	F
Nekewis io (Gabb)	.....	.....	.....	R
Olivella mathewsonii Gabb	.....	.....	.....	R
Pachycrommium? clarki (Stewart)	C	.....	F	C
Proximitra? cretacea (Gabb)	.....	.....	.....	F
Scaphander cf. *S. (Mirascapha) costatus* (Gabb)	.....	R	.....	F
Sinum obliquum (Gabb)	.....	.....	.....	R
Surculites mathewsonii (Gabb)	R	.....	.....	.....
Tejonia lajollaensis (Stewart)	R	R	.....	.....
Tejonia moragai (Stewart)	F	.....	F	.....
Tejonia? multiangulata Vokes	R	.....	.....	C
Turricula coaequalis (Anderson & Hanna)	.....	.....	.....	R
Turritella uvasana etheringtoni Merriam	F	R	.....	A
SCAPHOPODA				
Dentalium stentor? Anderson & Hanna	.....	.....	.....	F
ANNELIDA				
Rotularia tejonense (Arnold)	.....	.....	.....	A
ANTHOZOA				
Flabellum sp.	.....	.....	.....	R

R=rare; F=few; C=common; A=abundant

shown by the occurence of growth series. All of the associated bivalves, some of which show growth series, consist of single valves. Undoubtedly, current activity was responsible for separation of the valves, as indicated by the presence of crossbeds and scattered pebbles in the lower part of unit E_2. Such current activity, however, did not abrade the specimens, therefore they were not transported very far. Some of the bivalve shells have boreholes, presumably the result of gastropod predation.

The nature and condition of the fossils in the lower part of unit E_2, therefore, are indicative of an assemblage, affected to some degree by water turbulence and slightly mixed with plant remains of presumable terrestrial origin. Eleven of the molluscan genera have modern-day representatives, based on the work by Keen (1963). These genera include *Acila, Brachidontes, Corbula, Crassatella, Gari, Glycymeris, Pitar, Tellina, Crepidula, Cylichnina,* and *Turritella*. Most of these live within a depth range of intertidal or near intertidal to 100 fathoms (183 m), with the exception of *Acila* and *Glycymeris* occur only at U.S.G.S. fossil locality 237, and they occur in abundance. The presence of abundant specimens of *Turritella* and common specimens of *Corbula* and *Cylichnina* also at this locality would seem to rule out depths as great as 700 fathoms (1280 m). It is likely, though, that this locality, which is from the lowermost fossil horizon of unit E_2, is from the deepest-water deposit of unit E_2.

The abundance of oysters and the paucity of other fauna in the upper part of unit E_2 clearly indicate an environment that was different from that during the deposition of the lower part of unit E_2. Keen (1963) reported modern *Ostrea* as living within a depth range of intertidal to 20 fathoms (37 m).

The observed upsection gradational sequence of offshore siltstone (unit E_1), gastropod- and bivalve-bearing sandstone (lower part of unit E_2), oyster-bearing sandstone (upper part of unit E_2), and nonmarine(?) arkose (unit E_3) indicates a nearshore, regressive sequence of sedimentation, followed by uplift and truncation before the deposition of the nonmarine Sespe Formation. This paleo-environmental interpretation is in keeping with that reported by Kriz (1947), Howell (1975a, 1975b), Nilsen and Clarke (1975), and Clarke and others (1975).

AGE AND PROBABLE CORRELATION

The faunal assemblage of unit E_2 is similar in taxonomic composition to the middle Eocene "Domengine" and "Transition" mega-invertebrate "Stages" of Clark and Vokes (1936). More refined similarities are with parts of the "Domengine Stage" *Turritella uvasana applinae* fauna, as well as to part of the "Transition Stage" *Ectinochilus supralicatus* fauna described by Givens (1974) for middle Eocene assemblages from the upper part of the Juncal Formation in the Pine Mountain region, about 37 km northwest of the lower Piru Creek area. The Pine Mountain region is the nearest area to the report area that has received detailed biostratigraphic analyses of strata of Eocene age. Other than the "Transition Stage" molluscan fauna recognized from the Pine Mountain area and now the lower Piru Creek area, only two sections are presently known to contain assemblages referable to this particular "Stage" (Givens, 1974).

Although many of the molluscan species from unit E_2 have extensive faunal ranges, nine species occur only in "Domengine" or "Domengine" and "Transition" "Stages" elsewhere on the Pacific Coast. According to Givens (1974), those that occur elsewhere only in the "Domengine Stage" are *Proximitra? cretacea, Tejonia lajollaensis,* and *Pelecyora gabbi.* In addition, *Turritella uvasana etheringtoni* occurs only in the "Domengine Stage" (Merriam, 1941). Those species that occur only in both "Stages" include *Cerithium cliffensis, Coalingodea tuberculiformis, Crassatella uvasana semidentata, Glycymeris (Glycymeris) rosecanyonensis,* and *Glyptoactis domenginica* (Givens, 1974). None of the remaining molluscan species of the assemblage is restricted to younger or older "Stages" of the Eocene, hence all the faunal components are compatible with the middle Eocene age determination.

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LATE PROTEROZOIC–EARLY CAMBRIAN BIOSTRATIGRAPHY, CORRELATION, AND PALEOENVIRONMENTS, WHITE–INYO FACIES, CALIFORNIA–NEVADA[1]

By GAIL J. WIGGETT[2]

ABSTRACT

The White–Inya facies cantains a succession af fassiliferaus strata crassing the late Proterazaic–Early Cambrian transitian. The base af the Cambrian is proposed as the harizan containing the earliest skeletal metazaan, a mallusc(?), *Wyattia.* Faur Early Cambrian faunal zanes are present: the Nan–Trilabite Zone, the zone af *Fallotaspis* cf. *F. tazemmourtensis,* the *Holmia–Nevadella* Zone, and the *Bonnia–Olenellus* Zane. The sectian was depasited in an epeiric sea under stable tectanic canditians. Law, middle, and high intertidal, and semi–sheltered and restricted shallaw subtidal depasits accur in a masaic of laterally and vertically shifting environments. Law faunal diversity in the lawer part af the sectian is attributed ta unfavarable enviranmental canditians.

INTRODUCTION

The White–Inyo Mountains in Mono and Inyo Counties, California, and areas belonging to the same facies in Nye and Esmeralda Counties, Nevada, have received increasing attention recently because of their succession of fossiliferous strata spanning the Proterozoic–Phanerozoic transition. These strata appear to encompass the critical period without major stratigraphic breaks, and have yielded a diverse fauna of trace and body fossils (Cloud and Nelson, 1966; Nelson and Durham, 1966; Langille, 1973; Firby and Durham, 1974; Wiggett, 1973, 1974).

Regional sedimentary characteristics indicate a mosaic of shifting intertidal and subtidal regimes that are recorded both laterally and vertically. Tidalites are especially important in approximately the lower third of the section, where faunal diversity is low. Thus, in the combination of changes in faunal composition and in paleoenvironment throughout the section, the area affords an opportunity for study of the interactions between evolutionary changes and environmental stresses in determining biostratigraphic succession, community structure, and taxonomic diversity in developing early metazoan associations.

BIOSTRATIGRAPHY

The first indications of metazoan activity appear in the upper part of the Wyman Formation, where the trace *Planolites* (Langille, pers. comm., 1974; Wiggett, 1974) and *Scolicia*–like trails of possible molluscan origin have been identified.

In the White–Inyo facies, the mollusc(?) *Wyattia reedensis* (Taylor, 1966), small, "tubular fossils" (Firby and Durham, 1974), and trace fossils are the only unequivocal fossils predating trilobites. No hyolithids, brachiopods, archaeocyathids, or other groups appear this low in the section. Cambrian students are increasingly favoring placement of the base of the Cambrian at some horizon below the first trilobites, recognizing that several characteristic Early Cambrian skeletalized groups appear before trilobites in many well–studied areas (Rozanov, 1967). Hence, the base of the Cambrian is proposed herein at the base of the lowest *Wyattia*–bearing horizon, near the top of the Reed Dolomite. The remainder of the Reed is probably Vendian in age, as the upper Wyman may also be, since there is no evidence of significant unconformity between the two formations.

Four Early Cambrian faunal zones are recognized in the White–Inyo facies (figure 1). The upper three contain trilobites and are discussed here. The lowest zone is termed the Non–Trilobite Zone (Wiggett, 1973), following terminology suggested in Great Britain for correlative, fossiliferous, pre–trilobite strata (Cowie and others, 1972). The zone begins at the base of the range–zone of *Wyattia,* and ends with the appearance of the first trilobite. This zone could possibly be named for *Wyattia,* but this fossil's apparent uniqueness to California makes it unsuitable as the name–bearer for a zone that must eventually be correlated worldwide.

The Non–Trilobite Zone is probably correlative with the Tommotian Stage of the Siberian Platform and with Tommotian equivalents in Europe, Britain, and elsewhere. Beds containing *Wyattia* and the "tubular fossils" of Firby and Durham (1974) may also be correlative with the Stirling Quartzite horizon in the Death Valley facies where Langille (1974) found small, shelly fossils.

Trace fossils in this zone are indicative of crawling molluscs, arthropods, polychaetes (?), and other "worm–like" groups. They include *Plagiogmus(?), Rusophycus* and *Cruziana* (Cloud and Nelson, 1966), *Planolites montanus,* flat, ribbon–like scolicoid (molluscan?) trails, *Diplocraterion, Skolithos,* and agglutinated micaceous worm tubes.

A pattern of three Early Cambrian trilobite zones is now widely recognized (Rozanov, 1967; Rozanov and others, 1969; Kobayashi, 1972; Cowie and others, 1972). The three trilobite zones of the White–Inyo facies, correlative to those recognized in the Sekwi Formation, Canadian Cordillera (Fritz, 1972), are, in ascending order: the zone of *Fallotaspis* cf. *F. tazemmourtensis,* the *Holmia–Nevadella* Zone, and the *Bonnia–Olenellus* Zone. They are based essentially on major trilobite faunas discussed by Nelson and Hupé (1964) and Nelson and Durham (1966).

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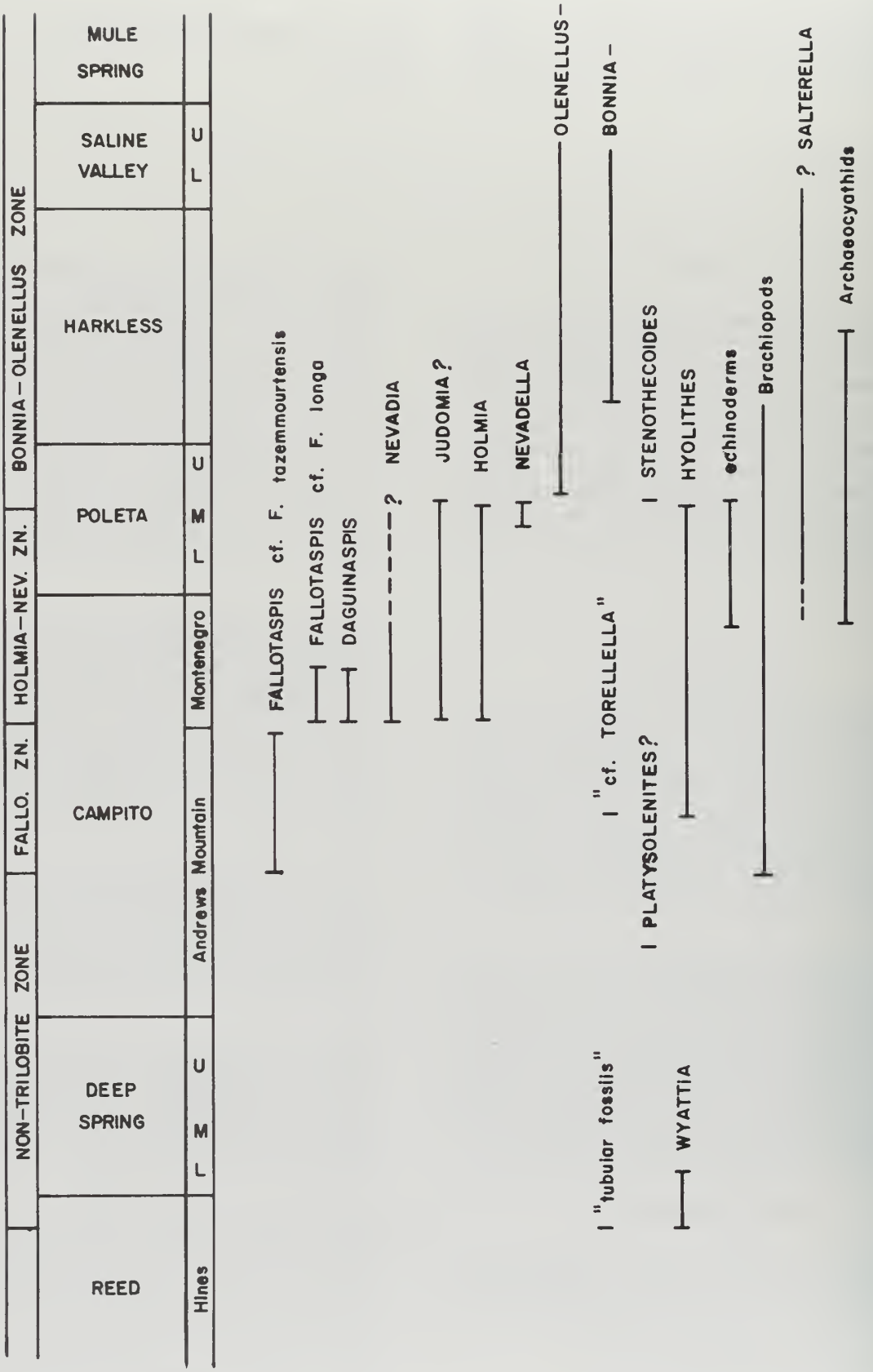

Figure 1. Founol zones ond ronges of body fossils discussed in text. Dato from references cited ond original observotions.

oldest trilobite zone includes the range–zone of *Fallotas-*
F. tazemmourtensis, but not (as defined here) that of *F.*
longa, Daguinaspis, or *Judomia(?).* Other characteristic
Cambrian body fossils appear for the first time in this zone
e 1). A new addition to the fauna is tentatively identified
hyolithelminth (?), *Torellella,* and it may extend the
ution of this Acado–Baltic and Siberian form into North
ica and into post–Tommotian beds. Trace fossils in this
again, represent arthropods, crawling molluscs, and poly-
s or other "worms". Additional molluscan traces and *Tei-*
nus rectus appear.

Holmia–Nevadella Zone is so named here in order to
nize the early appearance and long range of *Holmia* in the
nd the importance of *Nevadella* as a characteristic form
appears later in the zone. This zone is correlative with the
della zone of Fritz (1972). *Holmia* is widespread in correla-
orizons in much of the Atlantic province; hence it is useful
phasize its presence in the western North America portion
Pacific province. *Holmia* is not recorded, however, in the
Formation, although the related *Holmiella* is (Fritz,
. *Nevadella* occurs abundantly only in portions of the
Basin and the Cordillera.

base of the *Holmia–Nevadella* Zone is marked by the
rance of *Holmia,* in the lower part of the Montenegro
er of the Campito Formation. Here it occurs, near West-
Pass in the Blanco Mountain quadrangle, California, at a
y which has also yielded *Judomia?, Nevadia?, Daguinas-*
nd *Fallotaspis* cf. *F. longa,* in float from the same or closely
ated beds. Although the exact relationship between the
s of *F.* cf. *F. longa, Holmia,* and *F.* cf. *F. tazemourtensis*
ot been demonstrated, the latter has not been found above
ndrews Mountain Member of the Campito Formation.
(1972) included all occurrences of *Fallotaspis* and
inaspis in his *Fallotaspis* Zone, thereby incorporating into
one all of the Zones I through IV recognized in Morocco
upé (1952). The usage herein is more similar to that of
, in that *Fallotaspis* is allowed to range through more than
one.

he ranges of *Holmia* and *Fallotaspis* overlap, restricting
taspis to the earliest trilobite zone in the White–Inyo facies
nly be accomplished if the lower portions of the ranges of
mia?, and *Holmia* are also included in the lower zone. This
cts with the suggestion of Fritz (1972, p. 6) that the
alents of his *Nevadella* Zone were probably to be found in
rnia in beds bearing these genera and *Nevadella.* Further,
arison with worldwide stratigraphic data suggests that in-
n of *Holmia, Judomia?,* and *Nevadia* in the second zone is
appropriate.

esence of echinoderms and the non–pelecypod bivalved
usc, *Stenothecoides,* in the Middle Member of the Poleta
nation suggests possible correlation with the Sanashtykgol
on in the Siberian Altai–Sayan fold belt. The *Stenothec-*
is probably *S. knighti,* and it has previously been reported
from its type area near the Yukon–Alaska border (Yochel-
1969). The class Stenothecoida has never been reported
ously from California, and several undescribed steinkerns
enothecoides from the Silver Peak district of Nevada, men-
d by Yochelson (1969), are the only others reported from
Great Basin. Presumably they are not *S. knighti.*

rther correlation of the Poleta, with the middle member of
Wood Canyon Formation, Death Valley facies (Palmer, 1971), and with the Addy Quartzite, eastern Washington State, is suggested by presence of echinoderms, *Judomia?,* and *Nevadella* in the former, and *Nevadella* (Okulitch, 1951) in the latter.

Trace fossil diversity is high in the Montenegro Member and the Poleta Formation. Crawling molluscs are represented by *Plagiogmus?, Curvolithus,* and other scolicoid traces, and annelids or other "worms" by agglutinated micaceous tubes, *Planolites montanus* and other species of *Planolites,* and *Palaeophycus.* Probably two classes of arthropods are represented, by *Diplichnites* (attributed to trilobites) and by pellet–lined thalassinoid burrows. The latter are found in both the Poleta (Nations and Beus, 1974; Wiggett, 1974) and the Montenegro, and are the oldest known thalassinoid–type burrows. The next younger ones are found in Permian rocks. Burrows of this type are usually attributed to decapod crustaceans of an advanced order, but it seems probable that some early crustaceans may have developed a similar behavior pattern. Crustacean body fossils, however, are not known from Early Cambrian rocks, so that these traces suggest a possible range extension of the class. Star–shaped and radial feeding(?) traces also occur in the Montenegro.

The youngest trilobite zone is marked by the appearance of *Olenellus* in the uppermost Middle Poleta. Diversity of trilobites is very high in this zone, with as many as nine genera reported from one locality in the Saline Valley Formation (Palmer, 1964). Several new trace fossils appear, but overall trace fossil diversity is comparable to that in the *Holmia–Nevadella* Zone. Traces include *Bergaueria* (Alpert, 1973; Wiggett, 1974), *Skolithos linearis, Monocraterion* cf. *M. tentaculum, Planolites montanus, P. ballandus,* and *P.* cf. *P. ballandus, Bolonia lata, B.* new ichnosp., *Psammichnites,* and other scolicoid traces, star–shaped feeding(?) traces, and *Cruziana* cf. *C. fasciculata* and *Rusophycus* ichnosp. Both of the latter two belong to Seilacher's *Cruziana fasciculata* group (1970). Previously, the group has been found in Spain, Iran, Sweden, and Pakistan (Seilacher, 1970); the White–Inyo occurrences are the first in North America. This list of traces represents a composite of occurrences in both California and Nevada portions of the facies (Wiggett, 1974). Faunal groups represented include crawling molluscs, arthropods (trilobites), an increased diversity of annelids or other "worms", and burrowing anemones.

PALEOENVIRONMENTS AND FAUNAL DEVELOPMENT

The White–Inyo section was deposited in intertidal through shallow subtidal environments under conditions of tectonic stability, and with a shoreline of low gradient (Wiggett, 1974). Overall, sedimentation rates were generally low. Grain sizes range from mud to medium sand, with larger sizes rare, and the sediments are well sorted. Locally, rates of deposition of terrigenous material, amount of current activity, grain sizes, amount of circulation or restriction are variable, and units are somewhat heterogeneous. Lenticular deposits of terrigenous sediment and of algal limestone make the Wyman a very variable unit, while in the Montenegro, Poleta, and Harkless, archaeocyathids are found in carbonate lenses (McKee and Gangloff, 1969; Stewart, 1970; Wiggett, 1974). In tectonically stable conditions, such effects are probably due to shifting of laterally contiguous environments: migration of sand bars and shoals, buildup of mud

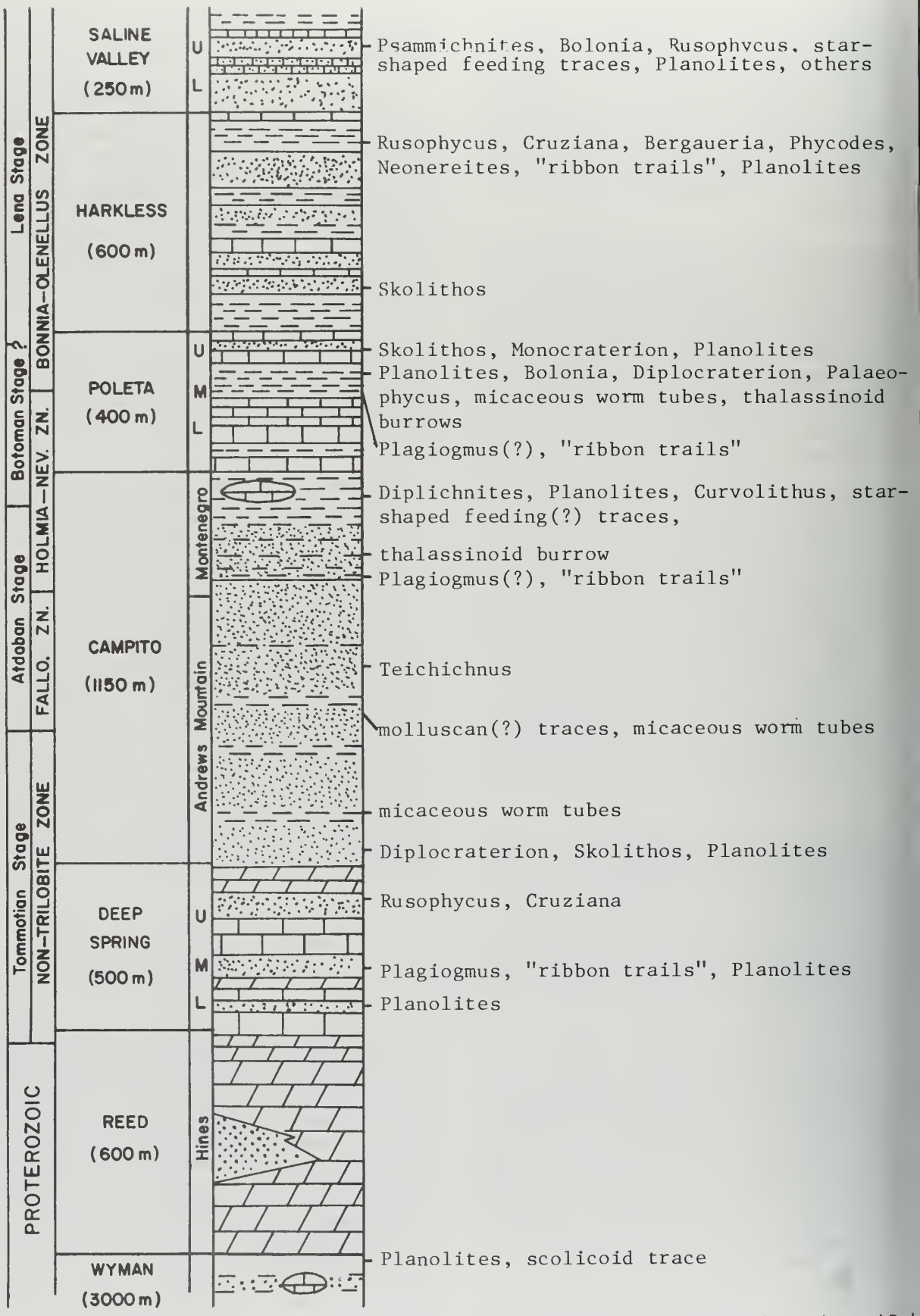

Figure 2. Stratigraphy and trace fossils, White–Inyo facies, California–Nevada. Stratigraphy slightly modified from Nelson and Durh (1966). Section is thinner in Nevada. Trace fossils from Wiggett (1974), with additions.

nds (with or without the aid of sediment–trapping by algae rchaeocyathids), and prograding of mud banks in the littoral (Roehl, 1967; Anderson, 1971). Sheltered lagoons or bays ably formed behind the mounds or shoals, and locally, re- sive conditions may have developed.

he lower third of the section was deposited mostly in very low water. Tidalites are important in the Reed and Deep ng Formations, but they also occur in the Wyman, Campito drews Mountain), and, higher in the section, in the Poleta Harkless. High, middle, and low intertidal deposits are ent (Wiggett, 1974). Indications of intertidal conditions are ndant, particularly in the Reed and Deep Spring, and include l cracks and mud crack casts, mud chips, current ripples, sand layers interspersed with clay laminae grading into r beds, and planar cross–bedded laminae. Algae and matolites, though not absolutely indicative of intertidal con- ns, are consistent with them, especially when associated so many other indicators. Stromatolites are found in the r Wyman, Reed, Deep Spring, and Poleta Formations ggett, 1974, and unpub. data, 1976).

he relative paucity of trace and body fossils in the lower part he White–Inyo section is attributed primarily to environmen- conditions unfavorable either to colonization of the substrate o preservation of the traces of colonization (Wiggett, 1974). y in the oldest part of the section, that is, in the lower Reed upper Wyman Formations, is lack of evolutionary progress bably a major factor. In particular, this is probably the case he upper Wyman, where only two traces and no body fossils known, but where environmental heterogeneity was great ugh that a variety of habitats was probably available for some d of exploitation.

n those parts of the section which are correlated with the mmotian Stage, environmental restriction was probably the or limiting factor. Evolutionary progress by that time had n sufficient to produce a varied fauna in, for example, the rian Platform and the Acado–Baltic area, and some groups are found in these areas appear later in the White–Inyo es. Faunal affinities with Morocco (Nelson and Hupé, 1964), Canadian Cordillera (Fritz, 1972), and Siberia area seen in older trilobites, the archaeocyathids (McKee and Gangloff, 9), and in the presence of *Platysolenites?* and cf. *Torellella*. s, if zoogeographic isolation was a factor earlier, it was no er important by *Fallotaspis* time. If the area were isolated ng the Tommotian, then a fauna of indigenous forms would xpected to characterize the Non–Trilobite Zone, given favor- environments and a few organisms. The poorly understood *attia* by itself makes a weak argument for isolation.

n the other hand, the modern intertidal zone is a rigorous ironment for invertebrates. Sparse bioturbation is character- of the high intertidal, and zones where currents vary from k water to high flow regimes. Organisms must contend with sure, erosion or burial, and fluctuating rates of deposition. noteworthy that those traces found in the Deep Spring are stly "facies–transgressive" forms, made by organisms that arently were able to adapt to a variety of conditions. They are products of vagile grazers or sediment–ingesters. When elling traces appear *(Skolithos, Diplocraterion)* it is in sedi- nts of more uniform grain size and less variable environment, he low intertidal and subtidal.

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TE PRECAMBRIAN DEPOSITIONAL ENVIRONMENT F THE PAHRUMP GROUP, PANAMINT MOUNTAINS, CALIFORNIA[1]

THEODORE C. LABOTKA[2] and ARDEN L. ALBEE[2]

ABSTRACT

The nature af the late Precambrian Pahrump Graup exposed in central Panamint Mauntains is indicative af a tectanically active viranment af depositian. The Crystal Spring Farmatian and Beck ring Dalamite were deposited an a shelf underlain by a platfarm older Precambrian gneiss. Part af the platfarm staad abave sea el (Warld Beater Island), and bath the Crystal Spring and Beck ring thin taward the island. At the end af Beck Spring depasitian western portian af the shelf sank while Warld Beater Island and jacent parts af the platfarm ta the narth emerged. The uplifted and pravided a local saurce far canglamerates in the lawer Kingsn Peak Farmatian which was depasited an the faundered shelf. gillite and quartzite filled the depressians, but Warld Beater Island ntinued to be a saurce far the averlying diamictite af the lawer ngstan Peak. The late Precambrian paleageagraphy af the Death lley area appears ta have been mare camplex than that suggestl by Wright and athers (1974).

INTRODUCTION

The late Precambrian Pahrump Group occupies a unique poon in the stratigraphic sequence of the Death Valley area, lifornia. These are the oldest deposits in the southwest Great sin which were not metamorphosed and strongly deformed or to late Paleozoic and Mesozoic tectonic events. The Pahmp Group marks the inception of grossly continuous sedimenion through the Paleozoic, and because the Pahrump is tensively exposed in the Panamint Mountains an opportunity provided to study the environment of the initial deposits in the rdilleran Geosyncline.

The Pahrump was divided into the three formations, Crystal Spring, Beck Spring Dolomite, and Kingston Peak, by Hewett (1940, 1956). This threefold division was recognized in the Panamint Mountains by Johnson (1957) and Albee and Lanphere (1962), and they correlated Murphy's (1932) formational names with those of Hewett (1956).

In the central Panamint Mountains (Telescope Peak quadrangle, figure 1) the Pahrump Group rests nonconformably on a complex of metasedimentary and metaigneous rocks approximately 1800 m.y. old intruded by porphyritic granite in the Pleasant Canyon area about 1400 m.y. ago (World Beater Complex; Lanphere and others, 1964). The Pahrump is overlain by Noonday Dolomite, Johnnie Formation and Stirling Quartzite; all lie below the lowest occurrence of lower Cambrian fossils in the Wood Canyon Formation (Diehl, 1974). A slight angular unconformity above the Pahrump Group is indicated by the truncation of the uppermost units in the Kingston Peak Formation by the Noonday Dolomite.

These rocks were regionally metamorphosed during the late Mesozoic (Lanphere, 1962) under low pressure conditions characterized by andalusite–bearing and cordierite–bearing assemblages in pelitic rocks. Isograds based on the stability of tremolite + calcite and diopside in calcareous rocks demonstrate east to west increase in metamorphic grade. Subsequently, the rocks were folded, and a north–northwest series of domes set en echelon by north trending faults was produced.

Wright and others (1974) outlined evidence for the existence of a west–northwest trough in the southern Death Valley–Kingston Range area (figure 1) from Crystal Spring time through at least Noonday Dolomite time and perhaps later. They call this trough the Amargosa Aulacogen and suggest that the west end opened into the north–northeast trending 'Beltian Geosyncline.' The central Panamint range is situated northwest of the Amargosa Aulacogen and within the postulated geosyncline, but the Pahrump exposed in the Panamint Mountains indicates a more complex paleogeography.

viewed by S. D. McDowell, R. F. Dymek, and R. E. Powell.

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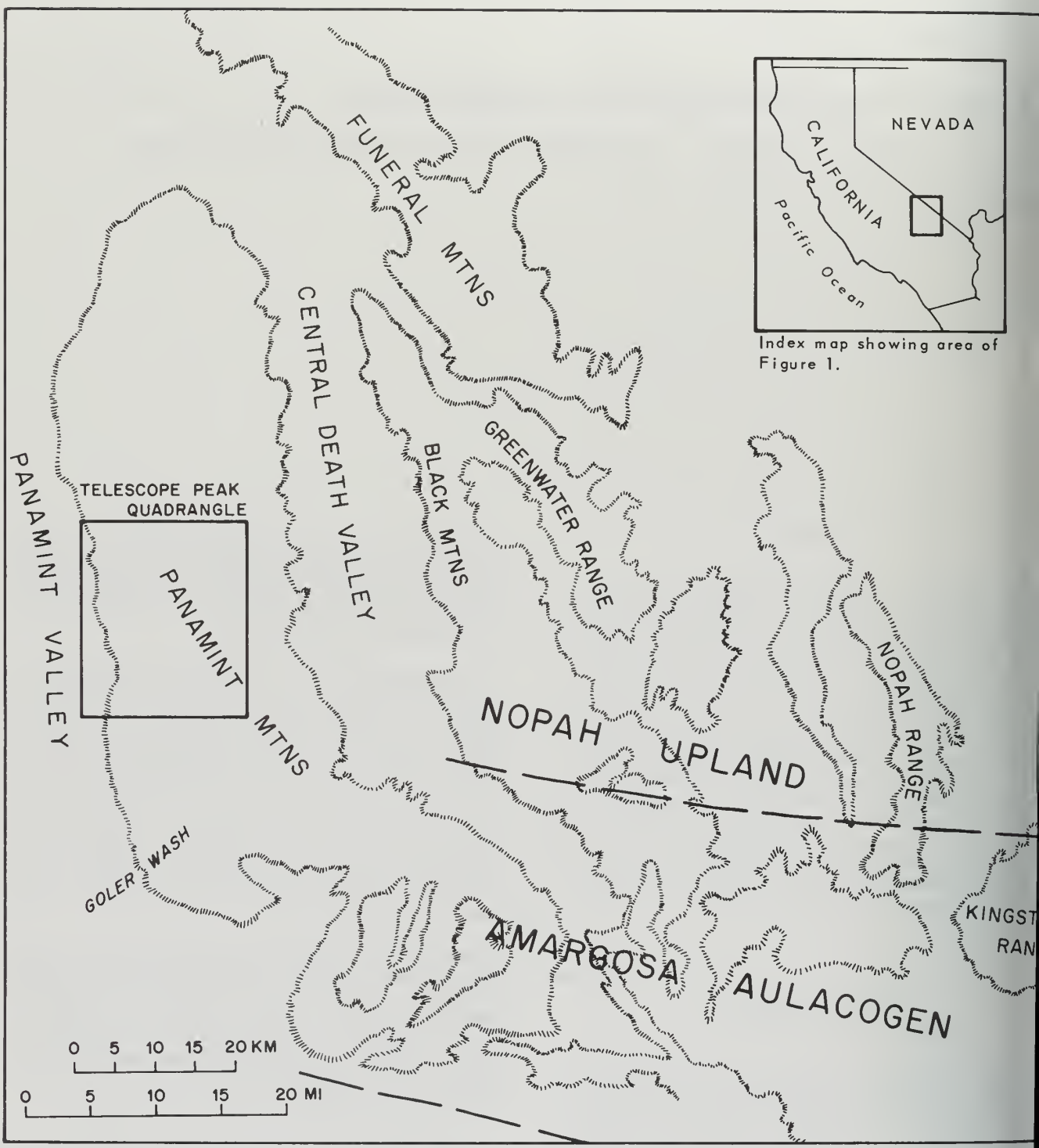

Figure 1. Location of the Telescope Peak quadrangle.

PAHRUMP GROUP IN TELESCOPE PEAK QUADRANGLE

tailed mapping which allowed the correlation of the Pana- Metamorphic Complex of Murphy (1932) with the Pah- Group and older Precambrian basement revealed a lex change in the lithologies within the Pahrump over short nces.

e superimposed Mesozoic metamorphism makes detailed graphic analysis difficult, but the metamorphic rocks do rve many primary sedimentary structures, particularly mil- er-scale bedding and graded bedding. The changes in the nesses and in the overall lithologies of the constituent units a first order determination of the paleogeography during ate Precambrian.

Crystal Spring Formation

e lowest formation in the Pahrump group is the Crystal g Formation, and throughout most of the Telescope Peak he Crystal Spring is considerably thinner than it is in the ern Death Valley area (Roberts, 1974). On the east flank e gneiss domes the Crystal Spring Formation is approxi- y 300 m thick, and west of the gneiss domes it thins to 100 m in Surprise Canyon, and it is absent south of Happy on.

st of the domes the Crystal Spring Formation consists of a quartzite-clast conglomerate, overlain by light grey dolo- and micaceous quartzite interbedded with argillite and nite. West of the gneiss domes the Crystal Spring section sts almost entirely of biotite-epidote-quartz schist, but a -brown weathering micaceous calcite marble occurs at the north of Hall Canyon. South of Happy Canyon, Crystal g is essentially absent, and clastic rocks of the Kingston Formation rest on early Precambrian basement. Locally, a quartzite-clast conglomerate of the Crystal Spring Forma- occurs at the base of the sedimentary section.

Beck Spring Dolomite

e Beck Spring Dolomite lies conformably on the Crystal g Formation throughout most of the Telescope Peak area. of the domes Beck Spring Dolomite consists of 200 to 300 massive-bedded dolomite and sandy dolomite. The upper Spring Dolomite on Sentinal Peak is dominantly clastic nterfingers with quartz-rich arenite. No continuous dolo- ed marks the top (Hewett, 1956, p. 26). Elsewhere on the side of the domes the contact between the massive Beck g Dolomite and overlying clastic rocks is sharp.

st of the gneiss domes the thickness of the dolomite varies about 180 m north of Surprise Canyon to 30 m just south rprise Canyon, and the Beck Spring Dolomite is absent of Happy Canyon. Nearly everywhere along the western in of the domes the Beck Spring Dolomite interfingers with ying clastic rocks. Between Surprise and Happy Canyons eck Spring Dolomite is greatly thinned and is extensively edded with clastic rocks. In Happy Canyon Beck Spring sts largely of dolomite breccia, and it rests on the older ment and pinches out. South of Happy Canyon Beck Spring mite is absent, and clastic rocks of the Kingston Peak For- on rest on older Precambrian basement.

In contrast to the gradational nature of the Beck Spring–Kingston Peak transition along the western margin of the gneiss domes, on top of the western gneiss dome in Surprise Canyon the contact between the Beck Spring Dolomite and the Kingston Peak Formation is an angular unconformity.

Kingston Peak Formation

The lithology of the Kingston Peak Formation varies considerably within the Death Valley area despite the ubiquitous conglomerates and pebbly mudstones. Lateral lithologic heterogeneity within the Kingston Peak in the Telescope Peak quadrange occurs with respect to distribution, size, and composition of clasts in conglomerates.

This lithologic heterogeneity is particularly acute in the lower Kingston Peak Formation (Surprise Member of Johnson, 1957), and the axis of gneiss domes separates an eastern and western "facies" of the Surprise Member. East of the domes the Surprise Member consists of thin-bedded argillite and arenite overlain by massive-bedded to unbedded conglomeratic greywacke, pebbly mudstone, and diamictite greater than 500 m thick. West of the domes the lower part of the Surprise Member consists of argillite, conglomerate, and arkose interbedded with dolomite and overlain by quartzite. The lower contact interfingers with Beck Spring Dolomite except above the western Surprise Dome and on the west flank of the World Beater Dome where the Surprise Member rests unconformably on Beck Spring Dolomite and older Precambrian basement, respectively. The conglomerates contain locally derived clasts of Beck Spring Dolomite and early Precambrian granite, and figure 2 shows the distribution of conglomerate and the lithology of the clasts. Arkose is a prominent constituent north of Hall Canyon, and it pinches out to the south, just north of Surprise Canyon. This argillite-dolomite-conglomerate sequence is overlain by platy quartzite, and the quartzite is overlain conformably by a diamictite similar to that east of the domes.

The Sourdough Limestone, a thinly banded, grey, micaceous limestone, about 20 m thick, separates the Surprise Member from the South Park Member of the Kingston Peak Formation (Johnson, 1957). The South Park Member consists of argillite and pelite overlain by cobble conglomerate and quartzite and is about 200 m thick.

PAHRUMP PALEOGEOGRAPHY IN THE CENTRAL PANAMINT MOUNTAINS

The most striking feature of the Pahrump group in the central Panamint Mountains is the stratigraphic change from east to west across the axis of the domes. The variation in thickness and lithology over short distances and the presence of locally derived clasts argue for the antiquity of the structural high now represented by domes and for tectonic activity during the deposition of the Pahrump Group.

Both the Crystal Spring Formation and Beck Spring Dolomite thin toward the World Beater Dome, and despite the erosion that occurred during the initial Kingston Peak deposition, the World Beater area must have stood above sea level during early Pahrump time (World Beater Island).

The floor of this part of the Pahrump basin is underlain by a platform of the older Precambrian basement. The local preserva-

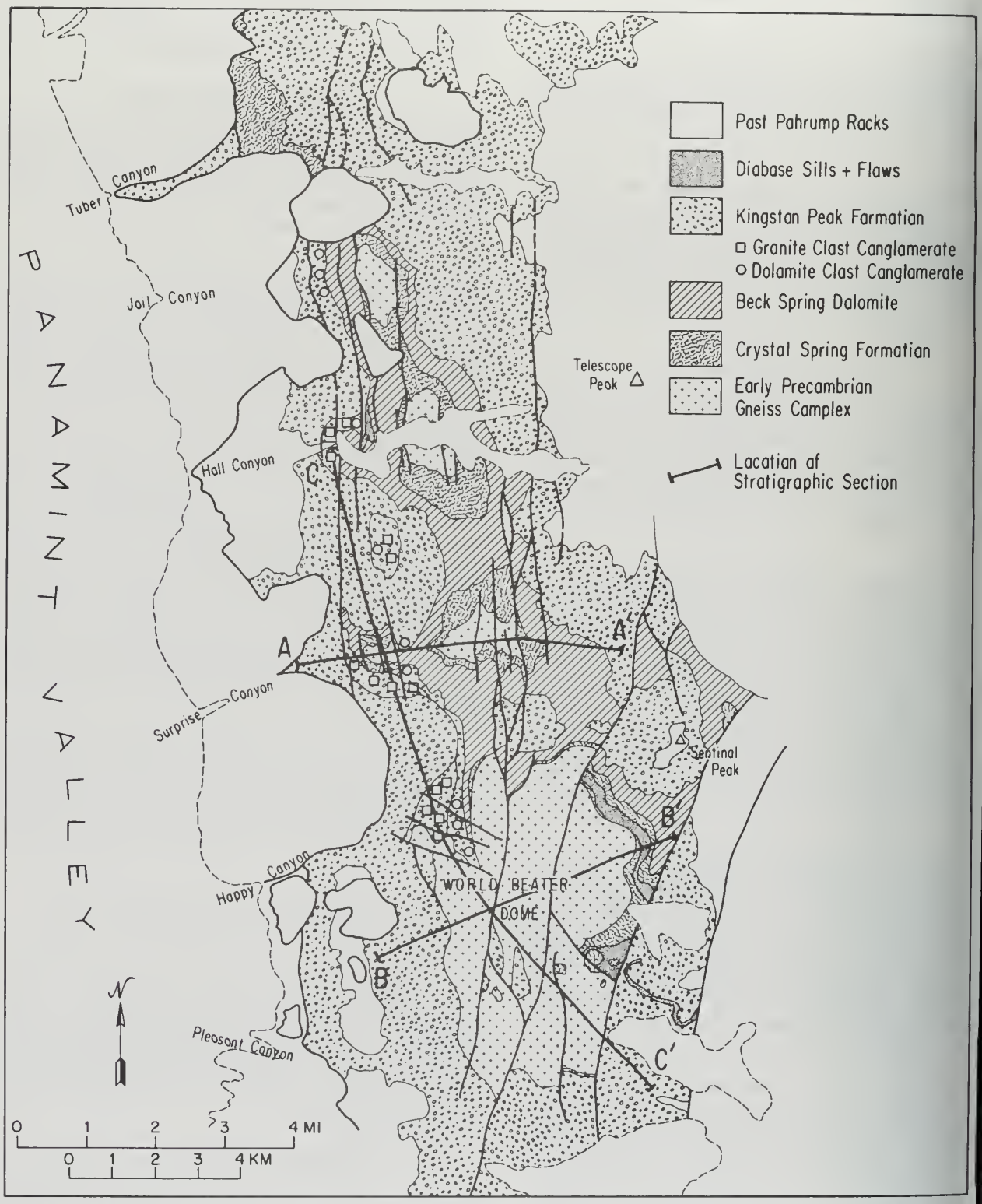

Figure 2. Generalized geology of the Telescope Peak quadrangle and location of stratigraphic cross sections in figure 3.

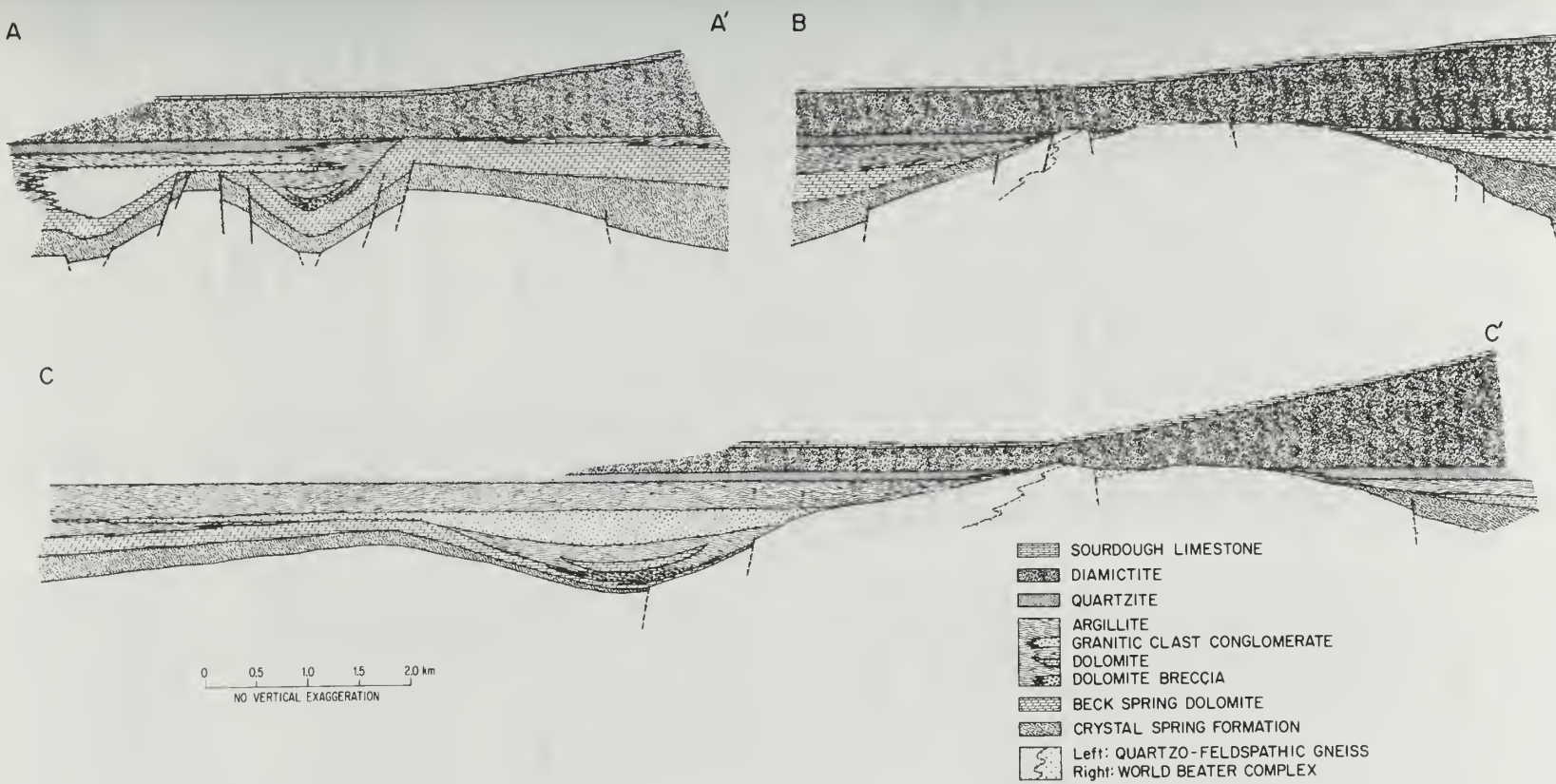

Figure 3. Generalized stratigraphic cross section of the Pahrump Group in Telescope Peak quadrangle. Sourdough Limestone occurs in the middle Kingston Peak Formation. Locations of the sections are given in figure 2.

tion of stromatolites in the Beck Spring Dolomite on Sentinal Peak and their widespread occurrence in the southern Death Valley area (Wright and others, 1974) suggest that the Beck Spring Dolomite represents a carbonate shelf developed on top of this platform.

The lower Kingston Peak Formation west of the gneiss domes is much different from that to the east, and figure 3 illustrates this change in lithology across the domes.

The intertonguing of Beck Spring Dolomite and clastic beds in the lower part of the Surprise Member west of the domes suggests that a portion of the carbonate shelf foundered. Simultaneously, World Beater Island and much of the platform to the north were uplifted and supplied both dolomite and granite clasts to the foundered area to the west and north. The lower Kingston Peak lacks conglomerate north of the western Surprise dome, and the unconformity developed on top of the dome indicates that the western Surprise dome was also a positive feature and a physical barrier to the dispersal of clasts derived from World Beater Island. Granitic clast conglomerates continued to fill the depressions and breached the top of the western Surprise dome. Subsequent to conglomerate deposition, argillite and pelite sedimentation continued, and these fine–grained sediments lapped over the conglomerate and World Beater Island. Deposition of the quartzite appears to represent the elimination of irregularities in the topography on the west side of the gneiss domes.

The first appearance of the unusual diamictites and associated rocks may represent a time line. Thus rocks east of the domes equivalent to the western argillite–conglomerate–quartzite sequence are thin and consist of argillite and pelite.

The peculiar texture and bedding characteristics of the Surprise Member are suggestive of glacial till (Johnson, 1957). A few thin, graded beds of argillite (often containing exotic "lonestones") do occur, and the presence of pillow structures in basaltic lava within the Surprise Member suggest that the Surprise Member is at least in part subaqueous and probably marine. The presence of Beck Spring Dolomite clasts up to 3 m in diameter and of clasts of the distinctive World Beater gneiss in the diamictite indicate that World Beater Island continued to be a source after the deposition of the quartzite in the western Surprise Member.

Intrusion of mafic sills occurred at least as late as earliest Kingston Peak time, and extrusive pillow lava flows occur in the lower Kingston Peak Formation. The presence of diabase clasts in the Kingston Peak below the lava flows indicate that igneous activity must have occurred over a significant period of time.

DISCUSSION

The geography and geology at the end of Beck Spring Dolomite deposition shown in figure 4 is largely schematic. Contours on the topography are arbitrary and are used only to show the interpretation of the geography.

It is suggested that the Pahrump Group in much of the Telescope Peak area was deposited on a shelf underlain by a platform of early Precambrian basement and that the platform stood above sea level in the World Beater Dome area. At the end of Beck Spring Dolomite deposition much of the western shelf foundered while World Beater Island and adjacent areas to the north were uplifted.

The limits of the Panamint Platform are imprecisely known. The Black Mountains appear to represent the eastern limit of Pahrump deposition north of the Amargosa trough (the Nopah Upland of Wright and others, 1974). A western positive area is indicated by the arkosic units in the western Surprise Member, and a western source for lower Crystal Spring conglomerates exposed south of Goler Wash was suggested by L. A. Wright (Pers. Comm.). Thus, another upland or another island may have existed west of the Panamint Mountains. The platform may extend as far south as Goler Wash in the southern Panamint Mountains where a relatively thin section of Kingston Peak Formation characteristic thickness of over 1000 m, the Beck Spring Dolomite is apparently replaced by a clastic limestone-argillite sequence, and the Kingston Peak Formation lacks diamictite and consists almost entirely of argillite and arenite. Mesozoic and Cenozoic deformation has obscured the transition from the platform sequence to the possibly deep water sequence north of Tuber Canyon, and this correlation is tentative.

The relation between the Panamint Platform and the Amargosa Aulacogen of Wright and others (1974) is not clear. If the platform extends as far south as Goler Wash, the mouth of the aulacogen may be partly blocked. Rocks which may represent deep water or geosynclinal environment do not occur south of Tuber Canyon, and the Amargosa trough does not open simply into a geosyncline.

Stewart (1972) argues that the initiation of the Cordilleran Geosyncline is defined by the development of a northwest paleoslope. If the opening of the Cordilleran ocean was accomplished by continental rifting, as proposed by many people (e.g. Stewart, 1972; Burchfiel and Davis, 1975; Gabrielse, 1972), the initial deposits are most likely to record a micro–environment dominated by local provenance and irregular basins of deposition. Only when the local environments are integrated can the overall northwest paleoslope by established. The Pahrump Group in Death Valley does record a complex late Precambrian paleogeography which strongly suggests the inception of continental rifting at the beginning of Pahrump time and continued rifting during middle Pahrump time.

ACKNOWLEDGMENTS

Geologic mapping in the Telescope Peak quadrangle by Marvin Lanphere and Douglas McDowell greatly contributed to the ideas proposed in this paper. Field work was supported by National Science Foundation Grant DES 75–03416 to A. L. Albee and by a Geological Society of America Penrose Bequest Research Grant to T. C. Labotka. The manuscript was critically reviewed by S. D. McDowell, R. F. Dymek, and R. E. Powell. This paper is Contribution No. 2809.

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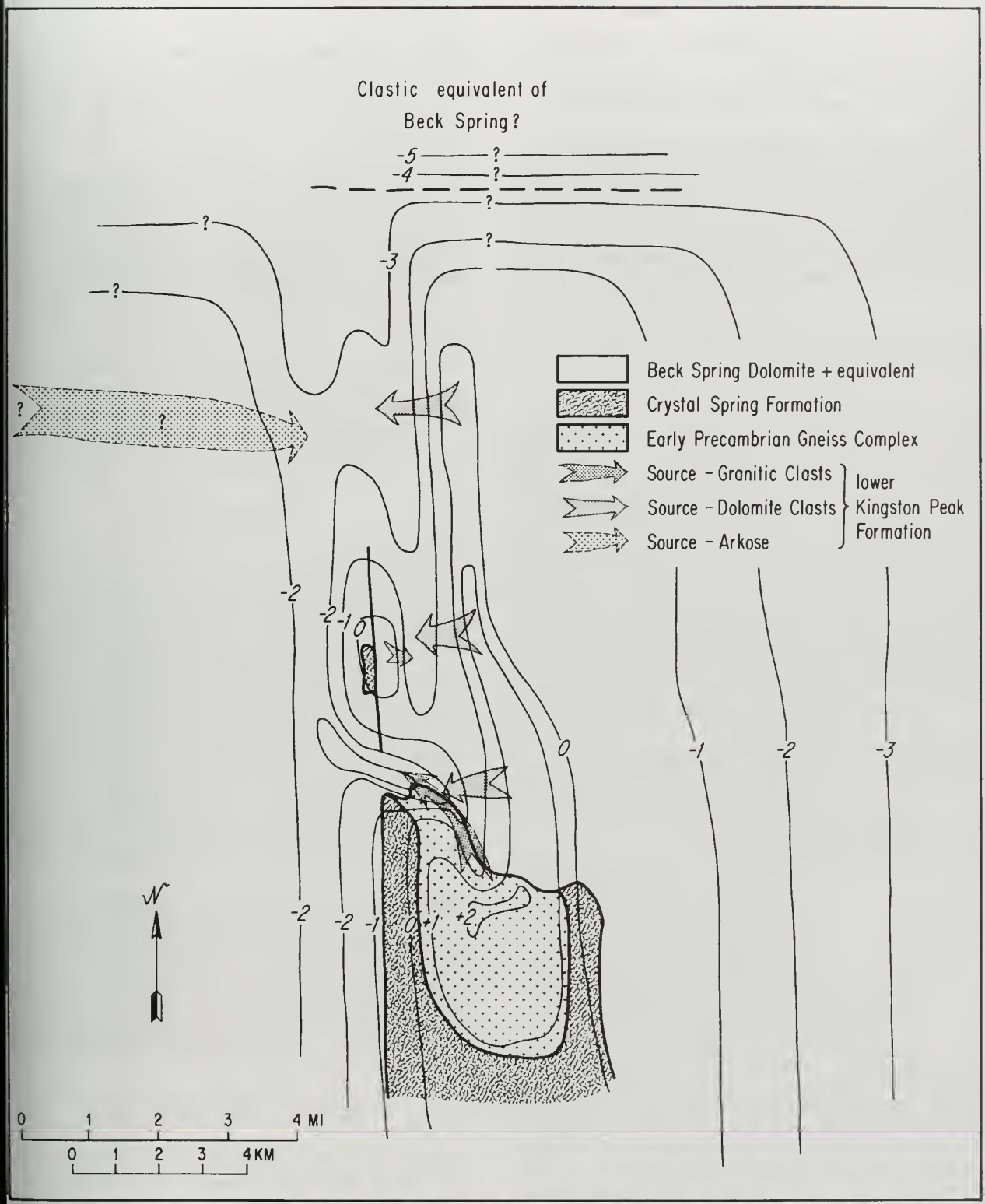

Figure 4. Postulated poleogeography ond poleogeology in the Telescope Peok quadrangle at the end of Beck Spring Dolomite deposition. Arrows indicate source of closts ond orkose in the lower Kingston Peok Formotion. Contours on the topography are orbitrary, but 0 represents sea level.

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AY MINERALOGY OF RECENT MARINE SEDIMENTS)UND ON THE)UTHERN CALIFORNIA OUTER CONTINENTAL SHELF[1]

RICHARD W. BERRY[2] and BRUCE NOCITA[2]

ABSTRACT

ay mineral analysis af the minus faur micran size fractian af 392 bax care les af recent sediment of the auter cantinental shelf af sauthern Califar- evealed a detrital suite af illite, chlarite, kaalinite, and expandable clay rals. Illite and expondable clays daminate the suite and were derived madern sails farmed under arid and semiarid canditians af weathering e cantiguaus cantinent and affshare islands. Chlarite is a minar canstitu- f the shelf sediments and is derived fram the same saurce as illite and ndable clay minerals. Kaalinite accurs as a minar canstituent and was d ta a saurce in the remnants af a Tertiary paleasal anshare. Differences an grain sizes af the canstituent clay mineral species allawed recagni- f the fallawing influences an their distributian: (1) Sauthern Califarnia ter Current, (2) bathymetric variatians and their effects an battam nts, and (3) wave–induced currents including affshare/anshare and hare components. Influence an clay mineral distributian patterns by tians within sediment saurce areas can alsa be detected.

INTRODUCTION

lay mineralogical analyses were performed on the minus four ron size fractions of the top 2 centimeters of 392 box cores n as a part of the *Bureau of Land Management Southern ifornia Baseline Studies and Analysis, Outer Continental lf Program.* In this context, the continental shelf includes the from shore to the Patton Escarpment. Box cores were taken R.V. Valero of the University of Southern California during fall 1975 and spring 1976. The program was coordinated by nce Applications, Inc., of La Jolla. Selected portions of the thern California Outer Shelf (hereafter referred to as High sity Areas) were sampled on a grid of approximately 2 meter intervals. Other portions of the shelf were sampled at gular intervals with core sites separated by as much as 20 meters. Figure 1 delineates areas of sampling activity and nguishes between high–density and low–density areas.

he purpose of this paper is to discuss the distribution of illite, rite, kaolinite, and expandable clay minerals in general terms to account for variations in their patterns of distribution. urate locations of sample sites as well as the semiquantitative tive percentages of the above named four major clay mineral stituents in the sediment at each sample site are in the ar- es of Geosecs of Scripps Institution of Oceanography, Uni- ity of California, San Diego, La Jolla.

ANALYTICAL PROCEDURES

he minus four micron size fraction of each sample was sepa- d at California State University, Northridge, as part of a icle size analysis of the total sediment sample. The clay–size tions of all samples were received by San Diego State Univer- as aqueous suspensions from which smears were prepared x–ray analysis. Ten percent of the samples were run in dupli- to test analytical precision.

ir–dried smears were subjected to x–ray analysis: (1) un- ted, (2) after remaining in an ethylene glycol desiccator for four days, (3) immediately after heat treatment to 350° C, and (4) after heat treatment to 550°C. Diagnostic peak areas were calculated as the product of the peak height over background and the peak width at one half the peak height. The following peaks were used to calculate approximate relative percent abundances of the four major clay mineral constituents:

illite—10 angstrams after glycolatian.
chlarite—14 angstrams after heating ta 550° C.
kaalinite—7 angstrams after heating ta 550° C.
tatal expandable clay minerals (smectitic + vermiculitic)—difference between 10 angstram peak areas after glycalatian and heating ta 550° C.

Chlorite and kaolinite calculations were based on the 7 and 14 angstrom peaks remaining after heating to 550° C because these same peaks after glycolation and after heating to 350° C both showed mutual interference from chlorite and kaolinite and were interfered with to varying degrees by a vermiculitic expandable constituent. Attempts to use peaks other than those at 7 and 14 angstroms failed owing to the low abundance and poor crystallinity of both chlorite and kaolinite. It should be clearly understood that the relative percent abundances in this report have little if any signifcance in terms of absolute abundances. The magnitude of increase or decrease in abundance from one sample to another for any one clay mineral can be shown by change in relative percent abundance. The warning is important because of the dangers inherent in overinterpretation of semiquantitative relative percentages such as those reported herein.

Twenty–nine replicate samples contained clay minerals in sufficient abundance to be above limits of detection. Results of the test of precision are tabulated below.

	Illite	Chlorite	Expandable clay	Kaolinite
Average difference in relative percent between replicates	3%	1%	3%	2%
Range of differences	0-10%	0-4%	0-9%	0-11%

RESULTS AND CONCLUSIONS

General Clay Mineralogy

Clay mineralogy of recent sediments of the Southern California Continental Shelf reflects the geology, climate and climatic history of the contiguous continental land mass and offshore islands. Clay minerals are detrital and show no evidence of diagenesis which is detectable by x–ray analysis. The clay mineral suite in the recent shelf sediment is essentially identical to a combination of those found in modern soils and paleosols on shore. The similarity extends beyond the four major clay mineral constituents and includes mixed layer clay minerals. The absolute abundances are also the same inasmuch as they can be compared.

Illite and expandable clays dominate the clay mineralogy of the shelf sediments; a normal clay mineral suite when the sediment source area is subjected to weathering under arid or semiarid conditions (Grim, 1968). Kaolinite occurs in small amounts (estimated 10 percent or less) throughout the entire shelf area

iewed by Edwin Hamilton and John Griffin.
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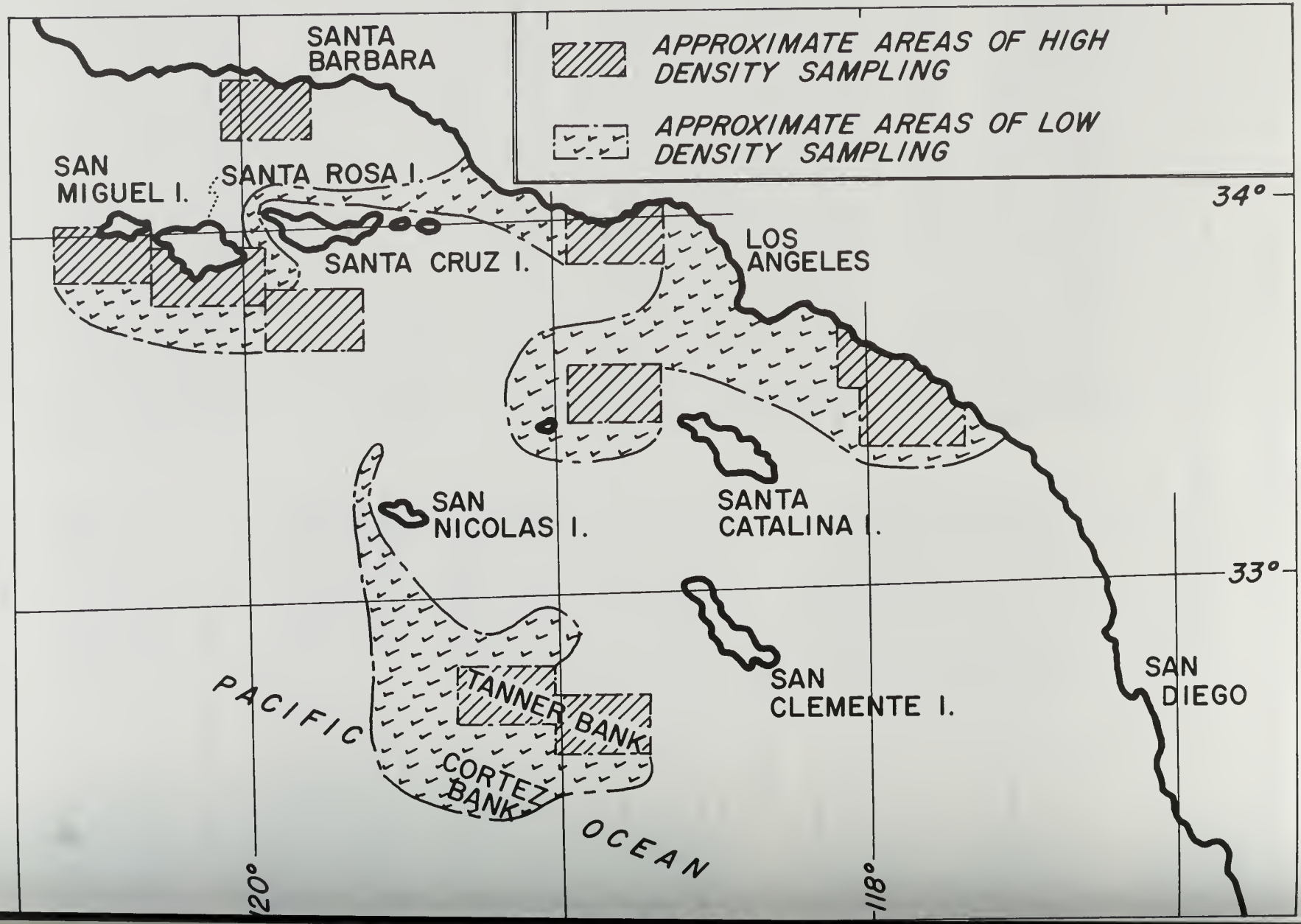

APPROXIMATE AREAS OF HIGH DENSITY SAMPLING
APPROXIMATE AREAS OF LOW DENSITY SAMPLING
SANTA BARBARA
SAN MIGUEL I.
SANTA ROSA I.
SANTA CRUZ I.
LOS ANGELES
SAN NICOLAS I.
SANTA CATALINA I.
SAN CLEMENTE I.
SAN DIEGO
TANNER BANK
CORTEZ BANK
PACIFIC OCEAN
34°
33°
120°
118°

udy. The source of kaolinite is a Tertiary paleosol, the rem- of which occur widely on the southern California main- (Abbott and others, 1976). The paleosol was formed under d climatic conditions and thus consists almost entirely of nite and quartz. Modern erosion continues to dissect rem- of the kaolinite–rich paleosol and transport small amounts aolinite mixed with illite and expandable clays of recent n to the Pacific Ocean.

e California shelf sediments contain chlorite of the same ty and abundance (estimated 10 percent or less) as is found odern soils of the contiguous continent with one exception. ments immediately to the west of Santa Catalina Island ain three or four times the amount of chlorite as is found where on the southern California shelf. The enhanced abun- e of chlorite results from the addition of a highly crystalline nesium–rich variety which occurs uniquely in the Santa lina Schist on Santa Catalina Island. The abundance of a Catalina Schist chlorite in the shelf sediment decreases distance from Santa Catalina Island until it disappears pletely several tens of kilometers west and north of the Is- .

xpandable clay minerals appear to consist of a physical mix- of smectite and vermiculite; but, because diagnostic peaks frequently masked by mixed layer clays and chlorite, it is er to refer to them as smectitic and vermiculitic expandable s rather than by their proper mineral names. X–ray data gest that there may be a transitional relationship between rite and vermiculitic expandable clay. There was no reliable to make even semiquantitative estimates of the relative pro- ions of smectitic and vermiculitic clay minerals. Hence, they e combined and reported as expandable clay minerals. Ex- dable clay mineral constituents of modern soils of the south- California mainland show the same characteristics as those cribed above for the shelf sediments.

Mixed layer clay minerals make up a significant proportion of clay–sized fraction of the sediment. The greatest majority of mixed layer clay minerals are random mixtures of chlorite illite, chlorite and expandable clay and possibly of illite and andable clay. Inability to measure the abundance of mixed r clays is very important to note because it is one of the sons why the percentage abundances of illite, chlorite, kaolin- and expandable clay minerals must be interpreted solely rela- to each other and *must not* under any circumstances be igned any absolute significance.

X–ray diffractograms of the clay sized fractions of both main- d soils and continental shelf sediments often show the devel- ment of an eleven angstrom peak after heating to 550° C. The ven angstrom peak is diagnostic of a random mixed layer clay neral consisting of 14 angstrom (chlorite) layers and 10 ang- om (expandable) collapsed layers and may reflect a mineral ase intermediate between chlorite and expandable clay miner- . Because the same mixed layer phase can be found both shore and onshore it is attributed to source area weathering cesses and not to diagenesis.

Non–clay Mineral Constituents

Crystalline clay mineral constituents of clay–sized fractions of ent continental shelf sediments show varying (and unmeasur- le) degrees of dilution by clay sized carbonate particles (prin- ally calcite) presumably of organic origin and by material hich is amorphous to x–rays (of unknown origin). In the anner–Cortez Bank area the dilution was excessive and re- ced the abundance of crystalline clay minerals to near or low the limits of detection by x–ray.

Clay Mineral Distribution

Clay minerals in shelf sediments were plotted and contoured on maps included in the final report to the Bureau of Land Management (Callahan, 1976). Salient features of the distribution of the four major clay mineral constituents are shown in a general way on figures 2 through 5.

Distribution of clay minerals in the benthic marine sediments of the Southern California Bight (figure 1) is controlled by the complex interaction of several factors: (1) variations in clay mineral grain size (within the clay–sized range of 4 microns to colloidal); (2) major ocean currents (the Southern California Counter Current in particular); (3) nearshore wave–controlled currents (including onshore/offshore components and longshore components); (4) influence on bottom currents by bathymetric irregularities such as shelves, slopes, and deep basins as well as channels and canyons which incise the slopes; and (5) variations in relative abundances of clay minerals in source areas.

In the offshore shelf regions of the Southern California Bight, the dominant control on clay mineral distribution is clearly the Southern California Counter Current. Less obvious but nonetheless clear control on clay mineral distribution is exerted by bathymetric variations wherever steep slopes are incised by canyons and channels which trend in directions that are nearly at right angles to the Southern California Counter Current (for example, west and northwest of Tanner and Cortez banks).

The very low relative abundance of kaolinite in the western region of Southern California Bight is attributed to a reduced abundance of kaolinite in the western source areas which is in turn attributed to the fact that this portion of source area was largely submerged under sea water during the time of kaolinite production on emergent land. A small but significant influence of variations in source area also can be seen in the sediments to the west and northwest of Santa Catalina Island where an unusually high abundance of chlorite is found. The relative abundance of chlorite diminishes to normal levels with increasing distance from the island. The abnormal abundance of chlorite is attributable to the Catalina Schist which contains an easily identifiable chlorite found only on Catalina Island. The occurrence of small amounts of the Catalina Schist chlorite in benthic sediments tens of kilometers to the west and northwest of Santa Catalina Island is clearly the result of transport by the Southern California Counter Current.

Sediments on the nearshore shelf and draped over the nearshore slope of the Southern California Bight are distributed under the influence of all controls mentioned previously. Although one control may dominate the distribution locally, there is no broad recognizably dominant control by any one factor.

Any one control on clay mineral distribution patterns may not affect the distribution of all four major clay mineral constituents equally. For example, local anomalies in the distribution of one or two clay minerals may correlate well with slope channels or submarine canyons while the remaining clay minerals may show no correlation whatever with channels in the same area. This is caused by variations in clay particle size within the clay size range. On the average, kaolinite and expandable clay minerals have a smaller grain size than illite or chlorite; however, it is quite possible that this size relationship will vary from locality to locality owing to poor crystallinity and attendant small grain size for illite and chlorite or to flocculation in situ and resultant enhanced grain size of the expandable clay minerals. If all the clay mineral types had a uniform grain size or if they exhibited a uniform variation in grain size, it would be reasonable to

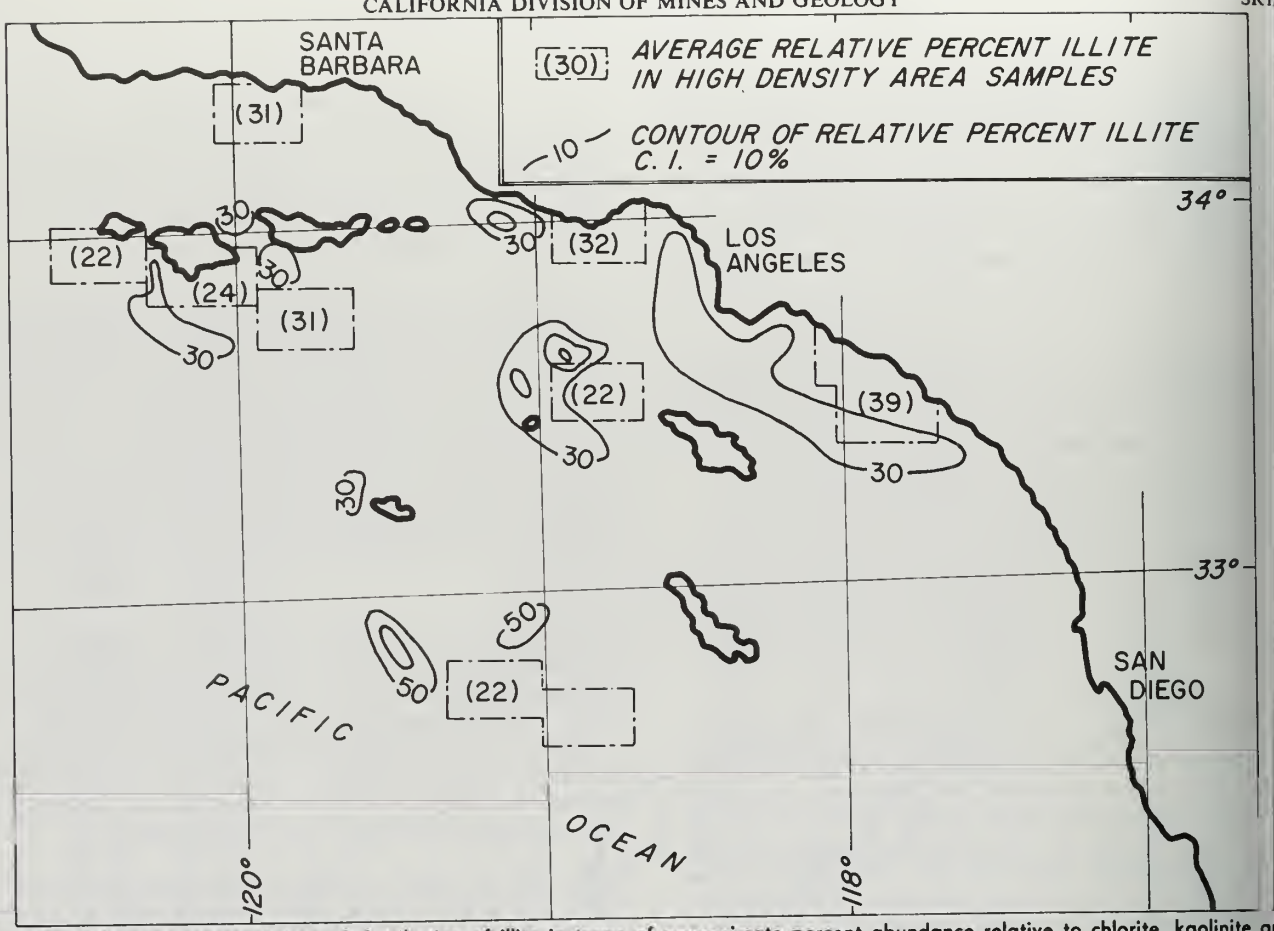


Figure 2. Schematic representation of distribution of illite in terms of approximate percent abundance relative to chlorite, kaolinite a expandable clay minerals. Only those contours are shown which are necessary to portray gross distribution patterns.

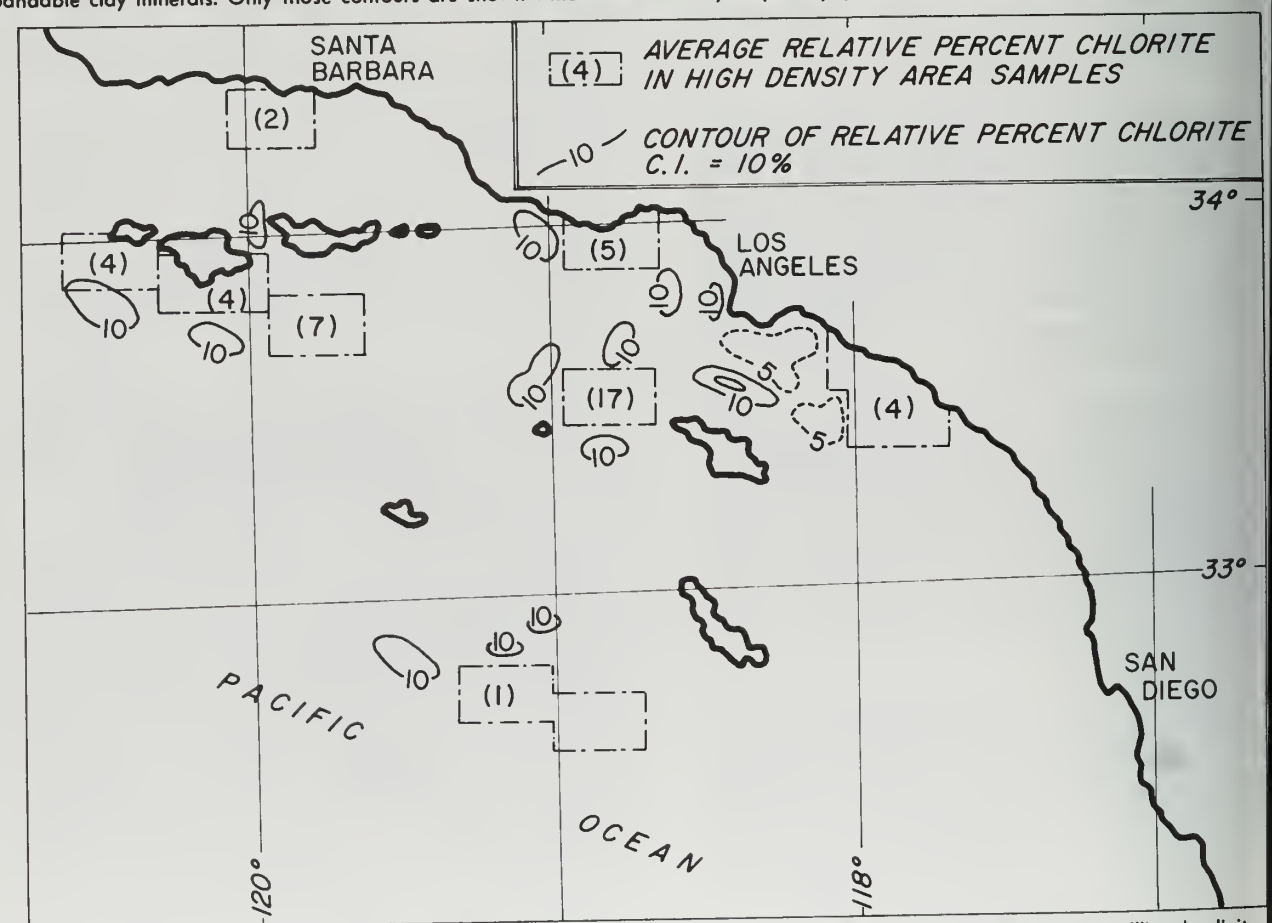
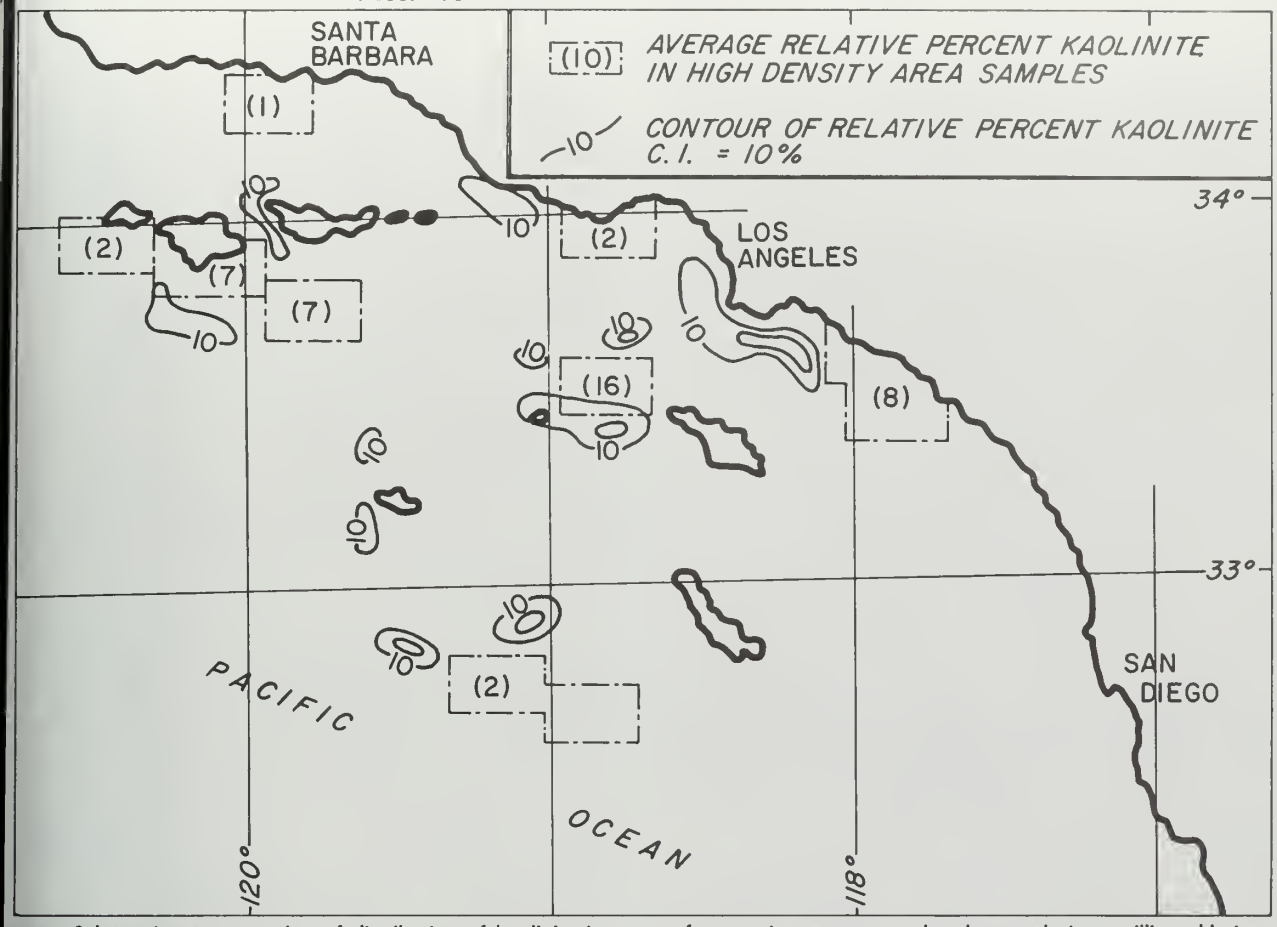
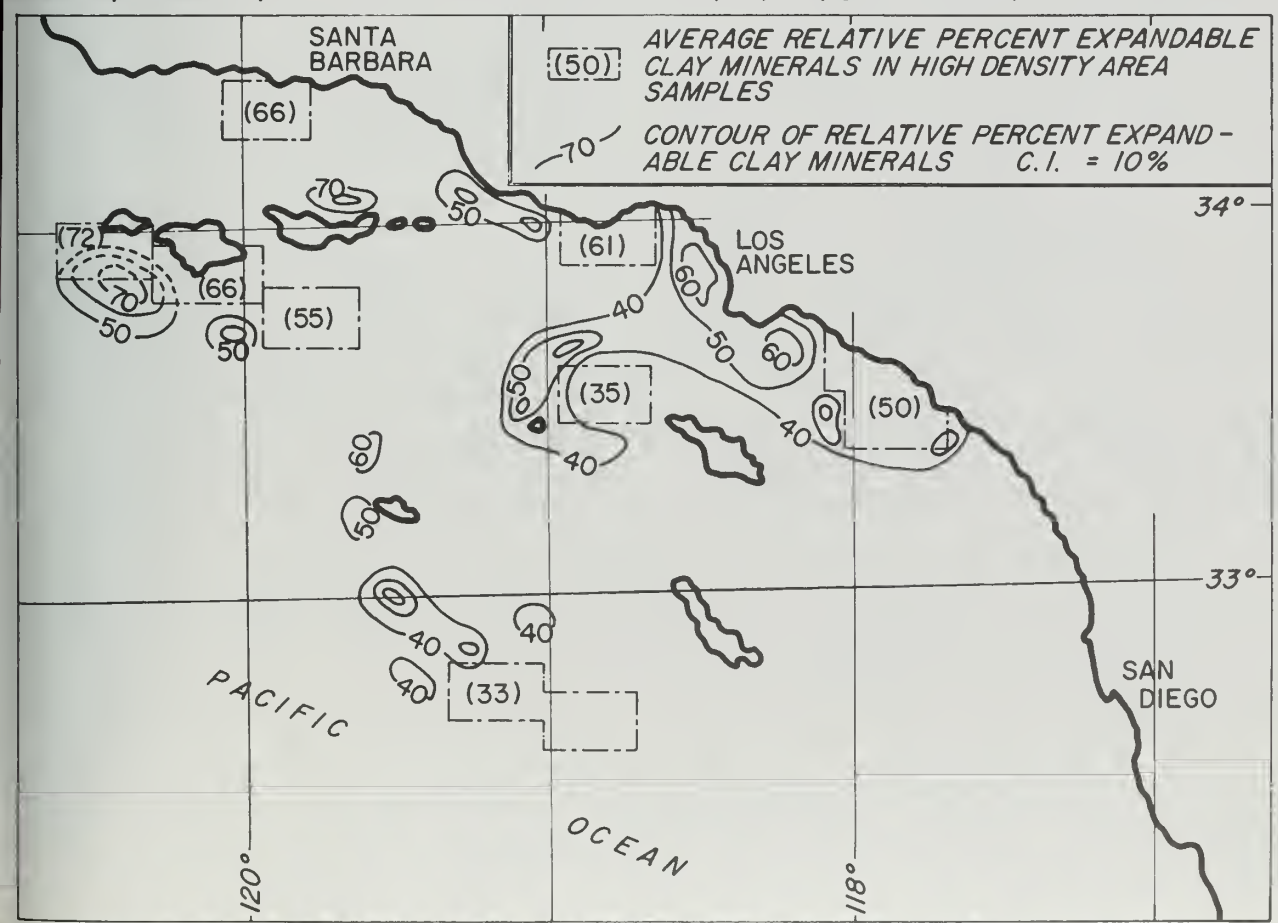


Figure 3. Schematic representatian af distribution of chlorite in terms of approximate percent abundance relative ta illite, kaalinite expandable clay minerals. Only those cantaurs are shown which are necessary to portray gross distributian patterns.



re 4. Schemotic representotion of distribution of koolinite in terms of opproximote percent obundonce relotive to illite, chlorite and andoble cloy minerols. Only those contours ore shown which ore necessory to portray gross distribution potterns.



ure 5. Schematic representation of distribution of expandable clay minerals in terms of approximate percent abundance relative to illite, lorite, and koolinite. Only those contours ore shown which ore necessory to portroy gross distribution potterns.

anticipate concordance of correlation of distribution patterns of all clay minerals with the dominant controlling factors. The lack of concordance indicates that there is little uniformity in grain size or grain size variations from one clay mineral to another.

Bioturbation is a potential source of noise in the data which must be considered. Hanor and Marshall (1971) show that organisms can churn the upper layer of sediment to depths of at least 25 cm. Until the degree of uniformity of clay mineralogy with depth can be determined, then the degree of noise introduced into the data by bioturbation cannot be assessed accurately.

SUMMARY OF RESULTS AND CONCLUSIONS

The clay–sized fraction of modern offshore shelf sediment of the Southern California Bight exhibits a suite of detrital clay minerals which includes illite and expandable clay minerals (both smectitic and vermiculitic) as principal constituents and kaolinite and chlorite as minor constituents. In the offshore regions, particularly in the vicinity of Cortez and Tanner Banks, the clay mineral suite is greatly diluted by clay–sized calcite (presumably organic) and by clay–sized material which is amorphous to x–rays (origin unknown). All samples contain mixed layer clays which consist of random mixtures of varying proportions of illite, chlorite, and expandable clay minerals.

In all respects the clay minerals of the recent shelf sediments are practically indistinguishable from their counterparts in the soils of the continental mainland and the offshore islands. No evidence of diagenesis can be discerned by x–ray. Thus the sources of the benthic clay minerals can be determined readily. Illite, chlorite and expandable clay minerals originate from erosion of modern soils which have developed under arid or semiarid climatic conditions. The kaolinite was formed during the early Tertiary under warm, humid climatic conditions. Post Early Eocene erosion has dissected the old kaolinite–rich soil (paleosol) but scattered remnants have been preserved and are now contributing sufficient kaolinite to the modern marine environment to account for the small amount found in the most recent benthic sediments. Mixed layer clays also occur in the modern soils and represent intermediate phases in the modern weathering process which ultimately reduces minerals of continental bedrock to illite and smectite.

The Southern California Counter Current exerts primary control on the distribution of clay minerals in recent sediments of the outer portions of the continental shelf. Clear, but less important, control is shown by slopes which separate banks and basins and particularly by channels and canyons which incise the slopes. The influence of bathymetric irregularities is felt because of their effect on the direction and strength of ocean bottom currents. Further controls by variations in bedrock and soils of source areas are seen in the vicinity of Santa Catalina Island (as a result of the Santa Catalina Schist) and in the regional distribution pattern of kaolinite (minimal abundances of kaolinite in shelf sediments in the western portion of the area of study corrrelating with fewer known occurrences of early Eocene paleosols in western source areas).

Distribution of clay minerals in sediments of the nearshore shelf is irregular and the result of control by several factors operating simultaneously. There is no dominant regional control on the nearshore shelf distribution but bathymetric features (shelf, slope, and channels), wave–related currents (including onshore/offshore and longshore components), coastal irregularities and variations in source area exert controls which can be recognized locally.

All factors (with the exception of source area variations) exert a control on the relative abundances of clay minerals by virtue of their relation to strength and direction of ocean currents. Illite and chlorite exhibit a mean grain size which is larger than that for kaolinite and expandable clay minerals thus they will usually be transported shorter distances by ocean current than kaolinite and expandable clays. Unfortunately, all clay minerals can be found as particles which fall anywhere within the clay–size range. Clay mineral grain size varies with the intensity of weathering regardless of the variety of clay mineral. Furthermore, expandable clays may flocculate in the marine environment, thus enhancing their grain size. The general correlation between grain size and clay mineral species allows recognition of the effects of controls on clay mineral distribution suggested in this paper. The fact that the correlation between grain size and clay mineral species is not sharp prevents the effects of controls on clay mineral distribution from being consistent or of equal clarity for all four of the major constituent clay minerals.

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